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MEMOIR 416

QUATERNARY GEOLOGY OF NEW BRUNSWICK

V.N. Rampton
R.C. Gauthier
J. Thibault
A.A. Seaman

1984

Canada

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R.J. Fulton

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Preface

The surficial materials lying upon bedrock play a vital role in shaping the human activity and economy of an area. Efficient use of the land will only be possible if adequate information is available concerning the terrain.

This report contains a description and explanation of the unconsolidated surficial materials of the Province of New Brunswick, and an account of the geological processes that have been active in shaping this area during Quaternary time.

Studies and compilations such as this are indispensable to a variety of human endeavours, such as land use planning, terrain classification (e.g., agricultural lands, forest lands and productivity), locating aggregate for building sites and highway construction, groundwater resource studies, locating waste disposal sites, and mineral exploration through drift prospecting. This report will serve these needs as well as help to focus additional studies to solve local problems and to meet local requirements.

Publication of this report and map has resulted from co-operation between New Brunswick Department of Natural Resources, Mineral Resources Division, and Geological Survey of Canada. Modern mapping and description of the surficial geology was started in the 1940s by the Geological Survey. In the past few years, the rate of mapping coverage was accelerated through field work contracted by the Geological Survey and by the New Brunswick Mineral Resources Division, the latter operating within the Canada-New Brunswick Minerals and Fuels Subsidiary Agreement. When the Agreement and funding ended in March 1982, the Geological Survey agreed to arrange for the preparation and publication of this report and the accompanying map.

R.A. Price
Director General
Geological Survey of Canada

OTTAWA

Préface

Les dépôts de surface reposant sur la roche en place sont à l'origine même de la forme que prendront les activités de l'être humain et l'économie d'une région. L'exploitation rationnelle des terres ne sera possible que si l'on dispose des renseignements nécessaires sur les terrains.

Ce rapport décrit et explique la nature des dépôts de surface non consolidés de la province du Nouveau-Brunswick et donne une explication des processus géologiques qui ont façonné cette région pendant le Quaternaire.

De telles études et compilations de données sont un élément indispensable de certaines activités de l'être humain, comme la planification de l'utilisation des terres, la classification des sols (par exemple, les terres agricoles, les zones forestières et la productivité), la localisation de dépôts d'agrégats utilisés comme site d'édifices ou pour la construction de voies publiques, les études des ressources en eaux souterraines, la détermination de sites d'élimination de déchets de l'exploration minérale par la prospection de dépôts glaciaires. Ce rapport se prête à ces fins et aidera à orienter les prochaines études vers les besoins et problèmes locaux.

La publication du rapport et de la carte est le fruit de la collaboration de la Division des ressources naturelles du ministère des Ressources naturelles du Nouveau-Brunswick et de la Commission géologique du Canada. C'est au cours des années 40 que la Commission géologique du Canada à commencé à préparer des cartes modernes et à décrire la géologie de surface des régions. Ces dernières années, les nouvelles cartes ont été préparées à un rythme accéléré, grâce aux travaux sur le terrain exécutés par la Commission géologique et la Division des ressources minérales du Nouveau-Brunswick dont les travaux sont financés dans le cadre de l'Entente auxiliaire Canada – Nouveau-Brunswick sur les minéraux et les combustibles. A l'expiration de l'entente et du financement, en mars 1982, la Commission géologique du Canada a accepté de prendre les dispositions nécessaires à la rédaction et à la publication du rapport de la carte qui l'accompagne.

Le directeur général de la
Commission géologique du Canada
R.A. Price

OTTAWA

CONTRIBUTION TO
CANADA-NEW BRUNSWICK MINERAL AND FUELS SUBSIDIARY AGREEMENT

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Geological contributions were provided by R.C. Gauthier, largely through contract (OSQ81-00025) with Geological Survey of Canada, and by J. Thibault and A.A. Seaman, as employees of New Brunswick Department of Natural Resources.

CONTENTS

1	Abstract/Résumé
3	Précis
5	Introduction
5	General statement
5	Approach and scope
5	Previous studies
6	Acknowledgments
6	Physiography, bedrock, and drainage
7	Physiographic history
9	Surficial materials
9	Pre-Quaternary
9	Bedrock
11	Wisconsinan and/or pre-Wisconsinan
11	Colluvium and glacial deposits
13	Wisconsinan
13	Morainal deposits
19	Glaciofluvial deposits
22	Late Wisconsinan and/or early Holocene
23	Marine deposits
28	Lacustrine deposits
30	Undifferentiated marine and lacustrine deposits
35	Holocene
35	Organic deposits
35	Alluvial deposits
36	Marine deposits
36	Eolian deposits
37	Quaternary
37	Colluvium
37	Quaternary history
37	Pre-Late Wisconsinan
37	Landscape development
37	Nonglacial intervals
37	Glacial intervals
41	Late Wisconsinan
42	Chignecto phase
44	Bantalor phase
49	Millville/Dungarvon phase
50	Plaster Rock and Chaleur phases
54	Madawaska Phase
55	Late Wisconsinan and early Holocene climate
56	Holocene
56	River regimes and alluviation
56	Coastal events
57	Terrestrial events
57	Tectonic events
57	Conclusions
59	References and selected bibliography
70	Appendix 1: Description of Physiographic Divisions and Subdivisions of New Brunswick

Illustrations

	Map 1594A. Surficial geology, New Brunswick
in pocket	1. Physiography of New Brunswick
6	2. Tectonostratigraphic zones and general geology of New Brunswick
9	3. Ternary diagram for different textured tills
10	4. Tor on Colonels Mountain in Northern Miramichi Highlands
10	5. Roches moutonnées along the west coast of Deer Island
11	6. Colluvium on slope southwest of Popple Depot
11	7. Nivation terraces on slopes along West Branch of South Nepisiguit River
12	8. Wedges of cobbles and boulders outlining sorted polygons in sandy till in Northern Miramichi Highlands
14	9. Texture of matrix of tills in northern New Brunswick
15	10. Location of multiple till localities in upper and middle Saint John River valley
16	11. Hummocky ablation moraine near Nigadoo Lake
16	12. Sunbury moraine
17	13. Ribbed morainic topography near McAdam
17	14. Ribbed morainic topography southeast of Serpentine Lake
18	15. DeGeer moraine near headwaters of Bartibog River

19	16.	Large multiple-crested esker near Magaguadavic Lake
20	17.	Eskers, meltwater channels, and ice-contact glaciofluvial deposits in Gulquac Mountains
in pocket	18.	Selected marine and lacustrine water level indicators in New Brunswick
in pocket	19.	Radiocarbon dates from New Brunswick
in pocket	20.	Moraines and speculative ice marginal positions in New Brunswick
28	21.	Exposure showing alluvium overlying thinly bedded brown rhythmites and varved clays
29	22.	Location of late Quaternary lakes in New Brunswick
30	23.	Thinly rhythmically bedded clays over till along Green River
31	24.	Deduced relative sea level curves for the Saint John and Edmundston areas
38	25.	Profile of Miramichi Highlands showing physics of hypothetical eastward-flowing glacier
39	26.	Dissection and drainage patterns around the northwest edge of the Central Plateau east of Sussex
in pocket	27.	Wisconsinan ice flow patterns in New Brunswick
40	28.	Selected pre-Late Wisconsinan glacial phenomena
43	29.	Paleogeography during Chignecto phase
in pocket	30.	Late Wisconsinan and early Holocene relative sea levels
44	31.	Paleogeography during maximum extent of Bantalor phase ice
46	32.	Deduced relative sea level curves for east of Prince Edward Island and for Chatham, New Brunswick
47	33.	Location and elevation of significant submarine and terrestrial features affecting Quaternary water bodies in New Brunswick and adjacent areas
49	34.	Paleogeography during maximum extent of Millville/Dungarvon phase ice
51	35.	Paleogeography during Plaster Rock and Chaleur phases
52	36.	Meltwater channels on valley walls west of Sisson Branch Reservoir
54	37.	Rhythmically bedded clay under peat along Green River
57	38.	Weathering pits in a tor
71	39.	Dissection and drainage pattern in siltstones adjacent to Tobique River

Tables

7	1.	Physiographic subdivisions of New Brunswick
24	2.	Radiocarbon dates for southern New Brunswick
32	3.	Radiocarbon dates for northern New Brunswick
42	4.	Summary of glacial flow patterns and relative sea level during the Late Wisconsinan

QUATERNARY GEOLOGY OF NEW BRUNSWICK

Abstract

New Brunswick's main physiographic divisions, which broadly parallel the geology and tectonostratigraphic divisions, are the Edmundston Highlands, Chaleur Uplands, New Brunswick Highlands, and New Brunswick Lowlands. A variety of Quaternary materials overlie weathered and glaciated bedrock throughout the province as depicted on Map 1594A.

Parts of the Northern Miramichi Highlands likely were glaciated only prior to the Wisconsinan as is suggested by the presence of well developed soils and periglacial features there. The Central Plateau of the Caledonian Highlands was probably glaciated in the Early to Middle Wisconsinan as is suggested by erosion patterns peripheral to the plateau.

The Late Wisconsinan in New Brunswick has been divided into five phases: (1) the early Chignecto phase during which ice flow in eastern New Brunswick was dominated by the Escuminac Ice Centre in the Gulf of Saint Lawrence; (2) the Bantalor and Sackville phases during which the Gaspereau Ice Centre on the New Brunswick Lowlands strongly influenced ice flow in eastern New Brunswick; (3) the Millville/Dungarvon phase during which ice flow in south-central New Brunswick was restricted to terrain adjacent to Saint John valley in western New Brunswick; (4) the Chaleur and Plaster Rock phases during which a glacier readvance in Baie des Chaleurs and the final deglaciation of northern New Brunswick occurred; and (5) the Madawaska phase during which Saint John River valley was occupied by a large lake or inland sea. A moraine near the maximum position of Bantalor ice at Saint John dates at 13.2 ka, and a diamicton deposited during the Chaleur phase dates at 12.4 ka. Deltas and other sea level indicators along the Bay of Fundy suggest a fluctuating relative water level of the DeGeer Sea during Late Wisconsinan deglaciation. East of Saint John, maximum submergence occurred during the Chignecto phase. Near Saint John, maximum submergence likely occurred during the early Bantalor phase, whereas west of Saint John, major deltas marking a second high relative sea level formed slightly later. Isostatic rebound at Saint John isolated a large inland sea in Saint John valley following 12.3 ka. Large glaciomarine deltas along Baie des Chaleurs were formed in the Goldthwait Sea after the Bantalor phase.

The Holocene has been primarily characterized by alluvial infilling and terracing along major valleys, submergence of coastlines, and development of major peatlands.

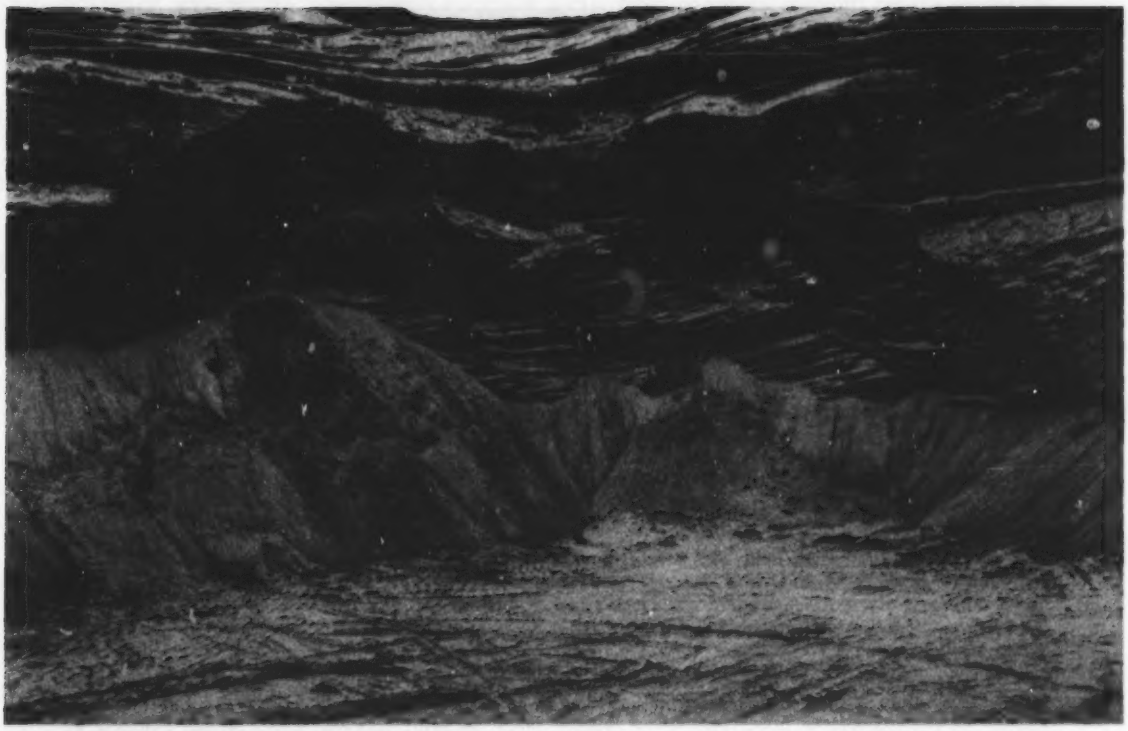
Résumé

Au Nouveau-Brunswick, les principales régions physiographiques sont les suivantes: les hautes-terres d'Edmundston, les bas-plateaux des Chaleurs, les hautes-terres du Nouveau-Brunswick et les basses-terres du Nouveau-Brunswick. De façon générale, ces zones correspondent aux formations géologiques de la province et à ses divisions tectonostratigraphiques. Comme l'indique la carte 1594A, divers matériaux quaternaires recouvrent une roche en place altérée et glaciée à la grandeur de la province.

Certaines parties du nord des hautes-terres Miramichi n'ont vraisemblablement été glaciées qu'avant le Wisconsin, ainsi que semble l'indiquer la présence de sols bien formés et de reliefs périglaciaires. Le plateau central des hautes-terres du Calédonien a probablement été glacié à un moment au cours du Wisconsinien inférieur ou moyen, comme semblent l'indiquer les structures d'érosion en périphérie du plateau.

Le Wisconsinien supérieur du Nouveau-Brunswick a été réparti en cinq phases: 1) le Chignectou inférieur au cours duquel l'écoulement glaciaire dans l'est du Nouveau-Brunswick a été dominé par le centre d'accumulation de glace Escuminac dans le golfe du Saint-Laurent; 2) les phases Bantalor et Sackville au cours desquelles le centre d'accumulation de glace Gaspereau sur les basses-terres du Nouveau-Brunswick a eu une forte influence sur l'écoulement glaciaire dans l'est du Nouveau-Brunswick; 3) la phase Millville/Dungarvon au cours de laquelle l'écoulement glaciaire dans le centre-sud du Nouveau-Brunswick a été limité au secteur adjacent à la vallée de la rivière Saint-Jean, dans l'ouest du Nouveau-Brunswick; 4) les phases des Chaleurs et Plaster Rock au cours desquelles ont eu lieu une récurrence glaciaire dans la baie des Chaleurs et la déglaciation définitive du nord du Nouveau-Brunswick; et 5) la phase Madawaska au cours de laquelle la vallée de la rivière Saint-Jean a été occupée par un vaste lac ou une mer intérieure. Une moraine formée près de la position d'avancée maximale des glaces de la phase Bantalor, à Saint-Jean, date de 13,2 ka et un diamicton mis en place au cours de la phase des Chaleurs remonte à 12,4 ka. La présence de deltas et d'autres indications de niveaux déjà atteints par la mer le long de la baie de Fundy semblent indiquer une fluctuation du niveau relatif de la mer De Geer au cours de la déglaciation survenue au Wisconsinien supérieur. À l'est de Saint-Jean, la submersion maximale s'est produite au cours de la phase Chignectou. Près de Saint-Jean, la submersion maximale a probablement eu lieu au début de la phase Tantalor, tandis qu'à l'ouest de Saint-Jean, d'importants deltas marquant un deuxième niveau relatif élevé de la mer se sont formés légèrement plus tard. Il y a 12,3 ka ou plus tard, un mouvement de remontée isostatique survenu à Saint-Jean a isolé une grande mer intérieure dans la vallée de la rivière Saint-Jean. De vastes deltas glaciomarins situés le long de la baie des Chaleurs se sont formés dans la mer Goldthwait après la phase Bantalor.

L'Holocène a été avant tout caractérisé par des apports alluviaux et la formation de terrasses le long des principales vallées, par la submersion des côtes et la formation d'importantes tourbières.



Frontispiece. *Stratified deltaic foresets over faulted ice-contact glaciofluvial sediments in Zealand delta (A.S.).*

Précis

Au Nouveau-Brunswick, les principales divisions physiographiques de la région des Appalaches sont les hautes-terres d'Edmundston, les bas-plateaux des Chaleurs, les hautes-terres du Nouveau-Brunswick (Miramichi, St. Croix et Calédonien) et les basses-terres du Nouveau-Brunswick. De façon générale, ces divisions s'apparentent aux divisions tectonostratigraphiques et à la géologie du Nouveau-Brunswick. Dans l'ensemble, le relief se présente sous forme d'une vaste surface d'érosion d'origine Tertiaire qui entoure des groupes isolés de collines. Par suite d'un soulèvement étendu, il y a eu morcellement du terrain et formation de vallées. Des cours d'eau ont été détournés à cause de la glaciation, du mouvement de la croûte et de la submersion survenus au Quaternaire.

On compte parmi les matériaux et les reliefs quaternaires représentés sur la carte 1594A: 1) le socle rocheux pré-Quaternaire, modelé par les glaciers et altéré; 2) des dépôts glaciaires (non différenciés), colluvions, régolithe et socle rocheux altéré du Wisconsinien et du pré-Wisconsinien, dans la partie nord des hautes-terres de Miramichi et le plateau central des hautes terres du Calédonien; 3) des dépôts morainiques et fluvio-glaciaires du Wisconsinien (pro-glaciaires et de contact); 4) des dépôts marins et lacustres du Wisconsinien supérieur et de l'Holocène inférieur; 5) des dépôts organiques, alluviaux, marins et éoliens de l'Holocène; et, 6) des colluvions du Quaternaire. Dans l'ensemble, l'épaisseur des dépôts non consolidés n'atteint pas 3 m. Les plus grandes épaisseurs se retrouvent seulement dans la moraine d'ablation, les dépôts fluvio-glaciaires, les plaines intertidales de l'Holocène et, par endroits, dans le till de fond. La plupart des dépôts morainiques se composent de till de fond caractérisé par des surfaces en pente indéfinissables, bien que de la moraine bosselée, ondulée et côtelée se retrouve partout dans la province.

Des pics, des reliefs périglaciaires et des sols très altérés et très oxydés caractérisent les régions élevées de la partie nord des hautes-terres de Miramichi et portent à croire que ces terres ont été recouvertes par les glaces seulement avant le Wisconsinien. Une zone fortement disséquée autour des flancs du plateau central des hautes-terres du Calédonien semble indiquer que cette partie a été recouverte par les glaces avant le Wisconsinien supérieur. Des stries relevées dans la région de Bathurst et dans le sud du Nouveau-Brunswick, datent peut-être d'avant le Wisconsinien, mais un réseau de stries orienté vers l'est et situé dans le sud du Nouveau-Brunswick date probablement du Wisconsinien inférieur étant donné que les stries se manifestent sur une plate-forme d'abrasion marine du Sangamon. Dans la vallée de la rivière Saint-Jean, un till que caractérise un horizon superficiel altéré date probablement d'une période antérieure au Wisconsinien; d'autres tills et d'autres dépôts gisant dans le sous-sol de cette région sont probablement liés à des phénomènes du Wisconsinien. Des matériaux interglaciaires ou interstadiaires reposent sous du till près de Sussex.

À la fin de la glaciation wisconsinienne, la glace s'est écoulée à partir de quatre centres principaux: 1) le centre d'accumulation de glace Escuminac, au large de la partie nord de l'Île-du-Prince-Édouard; 2) le centre d'accumulation de glace Gaspereau, dans les basses-terres du Nouveau-Brunswick; 3) la ligne de partage des glaces de la partie nord du Maine; et, 4) les centres d'accumulation de glace des monts Notre-Dame. De plus, les tracés de l'écoulement prouvent l'existence des <<champs de glace de l'est de Miramichi>> dans la partie est des hautes-terres de Miramichi. Il est probable que la glace du glacier des Laurentides soit descendue, au début du Wisconsinien supérieur, sur le nord-ouest du Nouveau-Brunswick. À mesure que se désintégrait le plus important centre d'accumulation de glace, de petits foyers d'écoulement glaciaire ont pris forme: il s'agit principalement du centre d'accumulation de glace du détroit de Northumberland, au nord de la péninsule Tormentine, et des centres d'accumulation St-Quentin et Sisson Branch, dans le nord-ouest du Nouveau-Brunswick.

Au cours de la phase Chignecto, survenue croit-on il y a 18 à 15 ka, le centre d'accumulation de glace Escuminac a dominé l'écoulement glaciaire dans l'est du Nouveau-Brunswick. À partir de ce centre, la glace s'est écoulée en passant par la péninsule Tormentine pour se déverser à l'extrémité sud de la baie de Fundy, et vers le nord-ouest dans la péninsule Acadian où elle s'est fusionnée avec une langue glaciaire provenant de l'ouest. Au cours de cette phase, il y a eu formation de deltas glaciomarins adjacents à la baie de Fundy. Le niveau marin relatif a atteint un maximum d'au moins 70 m le long de la baie de Fundy peu de temps après la disparition complète de la glace dans la baie.

À la phase Chignecto a succédé la phase Bantalor, survenue entre il y a 13,4 et 12,7 ka. Au cours de cette dernière le centre d'accumulation de glace Gaspereau a exercé une grande influence sur l'écoulement glaciaire dans l'est du Nouveau-Brunswick. Dans l'ouest et le nord du Nouveau-Brunswick, la glace s'est écoulée à partir de la région de la ligne de partage des glaces de la partie nord du Maine et du réseau des centres d'accumulation de glace des monts Notre-Dame. Des deltas glaciomarins, des complexes fluvio-glaciaires, des moraines frontales et interlobaires, et des zones de moraines d'ablation ondulées ou côtelées définissent les nombreuses marges glaciaires dont la formation date du moment où les glaciers ont commencé leur recul. Un till fossilifère, vieux de 13,2 ka environ, permet d'établir l'âge d'une importante moraine située à Saint-Jean. D'autres marges glaciaires sont définies par des rivages et des deltas marins situés à des niveaux semblables. Il semble que les élévations maximum du niveau marin relatif se sont produits lorsque la marge du glacier se situait près de Saint-Jean, soit il y a environ 13,2 ka, et qu'elle jouxtait les deltas Blagdon et Pennfield, il y a de cela moins de 13 ka. Le terrain, que recouvrait encore des restes de glace vers la fin de la phase, porte l'empreinte de niveaux de submersion relativement peu élevés. La mer DeGeer a envahi la vallée

de la rivière Saint-Jean au fur et à mesure que la glace reculait vers le nord. Les basses-terres du Nouveau-Brunswick ont peut-être été recouvertes seulement d'eau douce, la topographie et les mouvements différentiels de la croûte empêchant la pénétration d'eau salée en provenance de la mer Goldthwait.

Après la désintégration du centre d'accumulation de glace Gaspereau, la glace s'est écoulée principalement dans les basses-terres contiguës à la vallée de la rivière Saint-Jean, dans l'ouest du Nouveau-Brunswick, et dans la baie des Chaleurs, dans le nord du Nouveau-Brunswick. Cette phase glaciaire, appelée << Millville/Dungarvon>>, s'est produite il y a 12,7 à 12,4 ka. Une importante zone de moraine côtelée a été mise en place le long de l'axe ouest de la partie sud des hautes-terres de Miramichi et des hautes-terres de St. Croix, et un important delta s'est formé dans la mer DeGeer près de Fredericton. De grandes quantités de matériaux fluvio-glaciaires de nature subaquatique ont été mises en place dans la vallée de la rivière Saint-Jean à mesure que la marge glaciaire régressait vers le nord.

Les phases Plaster Rock et Chaleur, survenues entre il y a 12,4 et 12 ka, sont associées à la dernière avancée glaciaire d'importance dans la baie des Chaleurs et la désintégration définitive de la ligne de partage des glaces de la partie nord du Maine et des centres d'accumulation de glace des monts Notre-Dame. Il semble qu'il y a eu récurrence dans la baie des Chaleurs au cours de la phase Chaleur; les coquilles trouvées à la base d'un diamicton associé à cette avancée datent, approximativement, d'il y a 12,4 ka. La formation du <<Dame Restigouche>>, importante moraine interlobaire, a eu lieu au cours du recul de la marge glaciaire. Des matériaux glaciomarins ont été mis en place dans la mer Golthwait à mesure que la marge glaciaire régressait vers les hautes-terres environnantes. Au début de la phase Plaster Rock, la marge du glacier à reculé vers l'est à partir des hautes-terres d'Edmunston, après une récurrence mineure. Au méme moment, la partie supérieure du réseau hydrographique de la rivière Tobique a été détournée dans le lac glaciaire Upsalquitch, qu'interdisait un barrage formé par la glace dans la baie des Chaleurs. La ligne de partage des glaces de la partie nord du Maine a continué son recul, phénomène éventuellement responsable de la déglaciation de la partie supérieure de la vallée de la rivière Saint-Jean et de la formation des centres d'accumulation de glace St-Quentin et Sisson Branch sur le plateau de Saint-Quentin. Les chutes Reversing à Saint-Jean, dont l'émersion de la mer date de moins de 12,3 ka, ont créé la mer intérieure Acadia, grand lac situé dans la vallée de la rivière Saint-Jean.

Après la disparition de la ligne de partage des glaces de la partie nord du Maine et des centres d'accumulation de glace des monts Notre-Dame, le lac Madawaska a occupé la vallée en amont de Grand-Sault jusqu'à il y a environ 10 ka, alors qu'un soulèvement isostatique a mis à découvert une grande partie du fond du lac et donné naissance aux chutes situées à Grand-Sault. La mer intérieure Acadia a existé pendant quelque temps encore en aval de Grand-Sault, étant donné que le minimum du niveau marin relatif n'a été atteint, à Saint-Jean, qu'il y a 8 ka.

Les événements marquants de l'Holocène comprennent un soulèvement isostatique continu dans la plus grande partie de la province; le remblaiement alluvial de la vallée de la Madawaska et de la partie supérieure de la vallée de la rivière Saint-Jean; le recouvrement alluvial et la formation de terrasses dans les dépôts lacustres et marins des vallées des rivières; la formation de tourbières, principalement dans le nord-est de la province; et, une faible activité éolienne. L'élévation du niveau marin et la subsidence de la région côtière ont entraîné la submersion du littoral, l'inondation des vallées, et l'enfouissement par des sédiments marins d'arbres et de tourbe vieux de moins de 4 ka.

INTRODUCTION

General Statement

New Brunswick forms the core of the Canadian Appalachians and has experienced a unique style of Quaternary glaciation, isostatic depression and rebound, and submergence and emergence due to its geographic location. This report presents a summary of the present state of knowledge concerning these phenomena. It is the first intensive review of the Quaternary geology of the entire province since the turn of the century. This synthesis is made possible mainly because of completion of surficial geology mapping at scales of 1:250 000 (Gadd, 1973; Rampton and Paradis, 1981a, b, c) and 1:100 000 (Gauthier, 1982), simultaneously with development of new concepts of Maritime glaciation, isostatic rebound, and sea level changes (cf. Grant, 1977, 1980a, b; Quinlan and Beaumont, 1981, 1982).

Approach and Scope

This report and Map 1594A result from field investigations, aerial photography interpretation, and a review of literature and maps concerning the Quaternary geology of New Brunswick and adjacent areas. In order to integrate the observations and interpretations of all authors, the senior author conducted field reconnaissance with N.R. Gadd and the junior authors in their prime areas of investigation during the summer and fall of 1981. Fredericton map area (21G) was reviewed with N.R. Gadd; northern New Brunswick comprising Edmundston, Campbellton, and Bathurst map areas (21N, O, P) with R.C. Gauthier; and the southern edge of the Woodstock map area (21J) and northeastern half of the Fredericton map area with A.A. Seaman and J. Thibault.

Maps and reports by the junior authors were reviewed as they became available following the field reconnaissance. The senior author completed an independent field reconnaissance of much of the province as an aid to integrating the various map units of the different authors and to understanding their interpretations of the Quaternary geology. The nature of the units used was determined and phenomena critical to the interpretation of the geological history were examined during this final reconnaissance. A final review of the aerial photography for the province was completed in order to: 1) redefine units as necessary to integrate the 1:250 000 maps of Gadd (1973) and Rampton and Paradis (1981a, b, c) and the 1:100 000 maps of Gauthier (1982), which form the basis of the final Quaternary geology map and 2) understand the distribution of phenomena relevant to patterns of deglaciation and submergence of the province.

Results of detailed geological studies during 1982 by N.R. Gadd in the Saint John area, by Rampton in the eastern part of the Clarendon Hills (Thomas et al., in press), by Seaman in the Blackville area, and by Thibault in the South Oromocto area are included in this report where appropriate.

A selected bibliography has been compiled that includes reports concerning matters of allied disciplines, such as archeology and soil science.

Previous Studies

Systematic observations and mapping of the Quaternary geology of New Brunswick was initiated by Matthew (1872, 1879, 1894a, c), and was completed by Chalmers (1881, 1882, 1883, 1885, 1886, 1887a, b, 1888, 1890, 1893, 1895, 1901, 1902), who published surficial geology maps for most of the province. "Chalmers' concept of local ice caps based on his many observations of striae is still basically valid with minor modifications: his maps of the surficial materials are very general and details from inaccessible areas are lacking, probably because aerial photographs were not available to him" (Rampton and Paradis, 1981a, p. 2). Ganong (1896a, b, c, 1897, 1898, 1899a, b, c, 1901, 1902, 1903, 1904, 1905, 1906, 1907, 1908, 1909, 1910, 1911, 1913) made numerous observations concerning the physical geography of New Brunswick.

The Quaternary geology of most of New Brunswick has been mapped at scales of 1:100 000 and 1:250 000. The New Brunswick part of the Fredericton (21G) map area was mapped by Gadd (1973), and of the Amherst (21H), Moncton (21I), Woodstock (21J), and Charlottetown (21L) areas by Rampton and Paradis (1981a, b, c) at 1:250 000. The Quaternary geology of the New Brunswick part of the Edmundston (21N), Campbellton (21O), Bathurst (21P), and Matane (22B) areas was mapped by Gauthier (1982) at 1:100 000. Recent detailed mapping of surficial geology in New Brunswick has mainly been along Saint John River valley (Lee, 1953, 1955, 1957, 1959a, 1962a; Kite, 1979), in areas of urban development (Melvin, 1966; McCutcheon and Thibault, 1981; Ruitenberg and Thibault, 1980a, b, 1981), and in conjunction with granular resource inventory (Hamilton, 1973a, b; Brinsmead, 1974a, b, c, d, 1977, 1978, 1979a, b; Kingston, 1974a, b; Lohse, 1974a, b, c, d, 1976, 1977; Hamilton and Carroll, 1975; Kingston and Lohse, 1975; Thibault et al., 1976; Finamore, 1977, 1978, 1979a, b, 1981; Brinsmead and Finamore, 1977; Finamore and Lohse, 1977; Barnett et al., 1977; Thibault, 1978b, c, 1979, 1980, 1981, 1982; Cooper, 1980; Seaman, 1981, 1982a; Seaman and Thibault, 1981; Hunter and Lavergne, 1982a, b; Lavergne and Hunter, 1982a, b).

Numerous studies have been completed dealing with special aspects of the Quaternary geology of New Brunswick. Most notable are those relatively recent studies relating to glacial history (Lee, 1953; Kranck, 1972; Grant and Prest, 1975; Grant, 1976a; Gauthier and Cormier, 1977; Mott and Foster, 1977; Gauthier, 1978a, 1979, 1980a, 1983; Legget, 1980; Seaman, 1982b), emergence and submergence of coastal areas (Grant, 1970, 1975b, 1977d, 1980a, b; Welstead, 1971, 1976; Scott and Medioli, 1980), submergence in middle Saint John valley (Kiewiet de Jonge, 1951; Kite, 1979; Gauthier and Thibault, 1980; Kite and Borns, 1980; Kite et al., 1982), weathering (Gauthier, 1980b; Wang et al., 1981; Veillette and Nixon, 1982); mass movement (Ruitenberg et al., 1976), karst phenomena (Schroeder and Arsenault, 1978), and paleoecology (Thomas et al., 1973; Mott, 1975b; Korpijaakko, 1976).

Soils investigations (Stobbe, 1940; Stobbe and Aalund, 1944; Aalund and Wicklund, 1949; Wicklund and Langmaid, 1953; Millette and Langmaid, 1964; Langmaid et al., 1964; Langmaid et al., 1976, 1980) have resulted in maps delineating the geography of parent materials (near-surface surficial materials). Detailed investigation of peatlands (Korpijaakko, 1975; New Brunswick Department of Natural Resources, 1978a-c, 1979a, b; Keys and Henderson, in press) have resulted in maps showing distribution of peatlands. Investigations concerned primarily with forest productivity (Van Groenewoud and Ruitenberg, 1982), till geochemistry (Abbott and Herbért, 1967; Thomas et al., in press), land-use planning (Hirvonen and Madill, 1978; Van Groenewoud and Ruitenberg, 1982), geohydrology (Carr, 1961a; Hobson and Carr, 1967), engineering geology (Prest and Lee, 1952; Lee, 1961, 1962b; Ruitenberg and McCutcheon, 1978), and archeology (Turnbull, 1977; Turnbull and Allen, 1978; Allen, 1981) detail certain aspects of the surficial geology. Observations regarding unconsolidated deposits and glacial history were often conducted during bedrock investigations (for a bibliography of publications concerning bedrock see Poole et al., 1970; Carroll, 1977; Ruitenberg and Fyffe, 1982; Ruitenberg and McCutcheon, 1982; Fyffe, 1982a).

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PHYSIOGRAPHY, BEDROCK AND DRAINAGE

New Brunswick is located within the Appalachian Region of Canada (Bostock, 1970). The main physiographic divisions of the province (Fig. 1) are broadly related to the tectonostratigraphic zones and structural elements of New Brunswick (Fig. 2) as outlined by Poole and Rodgers (1972), Ruitenberg et al. (1977), and Fyffe (1982a, b).

Bostock's (1970) physiographic divisions have been slightly revised to more closely parallel boundaries between physiographic elements and between tectonostratigraphic divisions (Fig. 1, 2). The most significant revisions are (1) subdivision of the New Brunswick Highlands into Miramichi, St. Croix, and Caledonian highlands, (2) extension of the Chaleur Uplands around the northern end of the New Brunswick Highlands (eastern part of the Jacquet Plateau and Chaleur Coastal Plain in Fig. 1), (3) defining the New Brunswick Lowlands as a subdivision of the Maritime Plain, and (4) inclusion of the area west and north of Edmundston (Saint-François Hills and southern part of the Madawaska Hills in Fig. 1) in the Edmundston Highlands, a newly defined subdivision of the Notre Dame Mountains (Table 1).

The Jacquet Plateau, here shown as part of Chaleur Uplands rather than the New Brunswick Highlands, has a plateau-like surface more similar to the surface of the northern Chaleur Uplands than to the hilly terrain of the northern Miramichi Highlands. In addition, the southern boundary of the Jacquet Plateau parallels the southeast edge of Tobique-Chaleur Synclinorium (Fig. 1, 2).

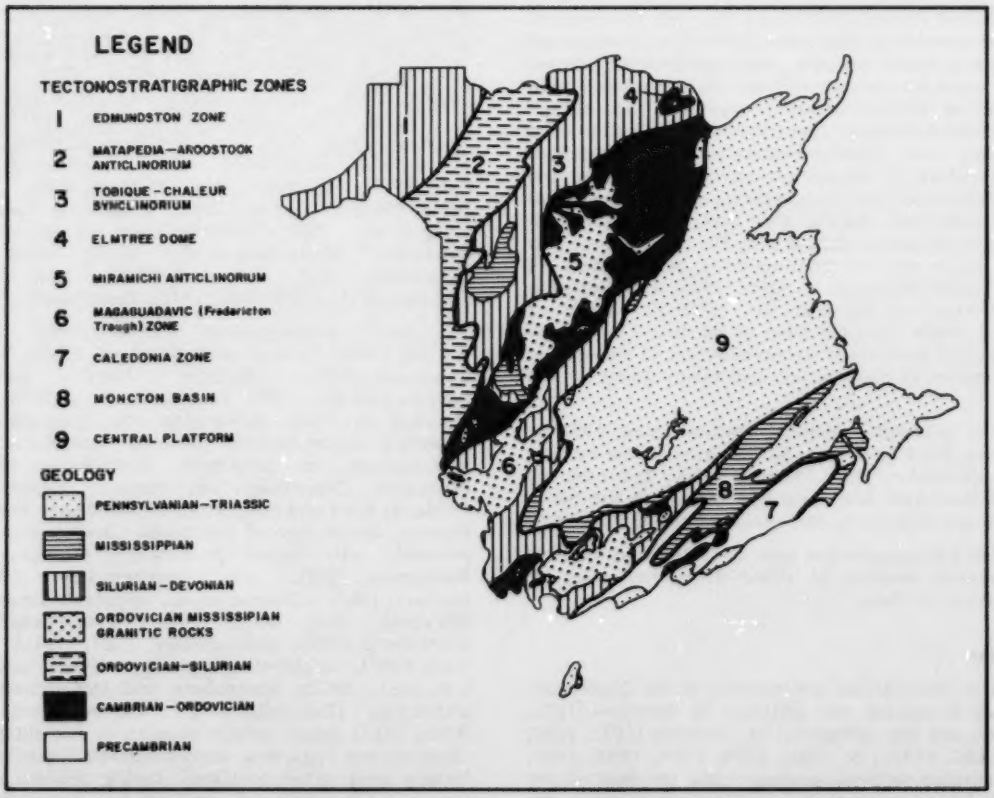


Figure 2. *Tectonostratigraphic zones and general geology of New Brunswick (modified from Poole and Rodgers, 1972; Ruitenberg et al., 1977; Potter et al., 1979; Fyffe, 1982a,b).*

Table 1. Physiographic subdivisions of New Brunswick

New Brunswick Division	Subdivision	Appalachian Region (Bostock, 1970)
New Brunswick Lowlands	Acadian Islands Acadian Peninsula Curventon-Bathurst Valley Central Lowland Tormentine Peninsula Grand Lake Basin Rusagonis Hills Nashwaak Hills	Maritime Plain
Caledonian Highlands	Anagance Ridges Moncton Basin Shepody Ridges Dorchester Upland Central Plateau Mispec Plateau	
St. Croix Highlands	Grand Manan Island West Isles Musquash Lowland Bocabec Hills Clarendon Hills Nerepis Highlands Scoullar Hills Magaguadavic Highlands Pokiok Hills	New Brunswick Highlands
Miramichi Highlands	Northern Miramichi Highlands Central Miramichi Highlands Southern Miramichi Highlands Eastern Miramichi Highlands West Bathurst Basin	
Chaleur Uplands	Williamstown Plateau Kintore Hills Tobique Basin Drummond Uplands Siegas Hills Central Saint John River Valley Saint-Quentin Plateau Jacquet Plateau Campbellton Hills Chaleur Coastal Plain	Chaleur Uplands
Edmundston Highlands	Kedgwick Highlands Madawaska Hills Saint-François Hills	Notre Dame Mountains

The Saint-François Hills and southern part of the Madawaska Hills have been removed from the Chaleur Uplands and incorporated into the Notre Dame Mountains (Edmundston Highlands) because their hilly nature is more similar to the Kedgwick Highlands and other subdivisions of the adjacent Notre Dame Mountains than to the plateau-like surface of the Chaleur Uplands to the east. Also, the Saint-François Hills and Madawaska Hills lie within the Edmundston Zone of the Gaspé Synclinorium (Poole and Rodgers, 1972) as does most of the southeastern part of the Notre Dame Mountains. The term Edmundston Highlands has been proposed for that portion of the Notre Dame Mountains encompassing northwestern New Brunswick and adjacent Quebec. Terrain within the Gaspé Synclinorium in the southeastern part of the Notre Dame Mountains is generally less rugged and mountaineous than the northwestern axis of the Notre Dame Mountains, which is within the Quebec Anticlinorial Zone (Poole and Rodgers, 1972). As such, the southeastern portion of the Notre Dame Mountains, which includes the Mégantic Hills, Edmundston Hills, southern Gaspésie, and other low hilly areas, might be considered as a highland distinct from the northwestern Notre Dame Mountains.

Detailed descriptions of the physiography, bedrock, and drainage of the physiographic divisions and subdivisions shown in Figure 1 and listed in Table 1 are given in Appendix 1.

Physiographic History

The basic configuration of the present New Brunswick landscape has evolved as a result of tectonic and erosional forces through the Cretaceous and Tertiary with modest modifications during the Quaternary. Older landscapes were covered by a flood of Carboniferous sediments, except possibly for the Central Highlands and Northern Miramichi

Highlands and the Notre Dame Mountains, which rise well above surrounding plateaus and are devoid of Carboniferous strata. They may well have continuously been land areas and subject to some degree of denudation during and subsequent to the Carboniferous.

Summits of the Notre Dame Mountains and the New Brunswick Highlands are presumed to be remnants of a large Cretaceous peneplain that extended throughout much of Atlantic Canada (Goldthwait, 1924). The Central Plateau of the Caledonian Highlands and the broad flat-crested ridges that characterize the deeply incised terrain of the Northern Miramichi Highlands adjacent to upper Nepisiguit River and the eastern part of the Kedgwick Highlands certainly define a high planar surface, which truncates strata of differing erodibility. The surface of the supposed Cretaceous peneplain is difficult to identify in other parts of the highlands where the topography is characterized by hills with crests at irregular elevation interspersed amongst broad valleys and basins.

Plateaus and lowlands that flank the higher plateaus and hills of the core of the New Brunswick Highlands appear to define an extensive erosion surface or pediment. A pediment is "formed by the parallel retreat of escarpments in areas of semi-arid climate. Basically, processes such as weathering, mass movement, local gullying, and basal sapping continuously erode the face of retreating escarpments. The material produced at the base of the escarpments is weathered and is transported across the pediment surface by creep and sheetwash. The expansion of the pediment continues regardless of the rock type in which the eroding escarpments are developed, although rock type undoubtedly affects the rate of retreat. Similar to stream erosion, pediments develop more rapidly in less resistant rocks such as shale and mudstone, and pediments in mountaineous areas generally are underlain by such rock types. The fact that the surfaces of pediments intercept various rock types tilted at different angles, however, indicates that pediments do not owe their existence solely to differential erosion of rock types or to structural control, but to the development of planar erosion by pediplanation or similar processes." (Rampton, 1982, p. 7). The surfaces of these plateaus and lowlands truncate or intercept various rock lithologies as exemplified by the Saint-Quentin and Jacquet plateaus and the New Brunswick Lowland and adjacent Eastern Miramichi Highlands. The erosion surface fringing the Miramichi Highlands, as defined by the New Brunswick Lowlands and Eastern Miramichi Highlands, is developed upon strata of varying erodibility such as Pennsylvanian sandstone and highly metamorphosed Ordovician volcanics and sediments (similar contrasts in rock types underlie other plateaus). The boundary between this erosion surface and the higher hills of the Miramichi Highlands is marked by an escarpment, in many places relatively sharp and in others, relatively subdued. The escarpment generally does not coincide exactly with fault lines or a marked contrast in rock lithologies and as such is probably a paleofeature of the pediplanation process.

Erosion and pediplanation will, of course, proceed rapidly in less competent rocks. Thus the core of the large pediment east of the Miramichi Highlands was formed on Pennsylvanian friable sandstones and soft shales. Pediments probably extended along stream valleys within the highlands. Pediments within the highlands were probably developed in most places on soft rocks. Typical is the basin in which Nepisiguit Lakes lie; it is underlain by Devonian sedimentary rocks of lesser competence than the Devonian volcanics and other more competent strata underlying the adjacent terrain (Davies, 1977). Most hills located on plateau surfaces, e.g. Blue Mountain north of Jacquet River and Blue Mountain east of Tobique River, are simply erosional residuals or monadnocks, characteristic features of pediplain surfaces.

Truncation of Triassic strata by the erosion surface or pediplain on the New Brunswick Lowlands and Musquash Lowland indicates that the surface postdates Triassic time. Goldthwait (1924) has assigned a Tertiary age to the development of his Atlantic Lowlands, which may be the correlative of the pediplain formed by the New Brunswick plateaus and lowlands.

Broad regional uplift affected New Brunswick during the late Cenozoic after formation of the pediplain. This uplift led to deep stream incision and dissection. Upper Nepisiguit River valley in the Miramichi Highlands and much of Saint John River valley probably formed during this interval. Locally, uplift does not appear to have been uniform and some block faulting may have occurred. For example, the markedly divergent elevations across Saint John Harbour of the Musquash Lowland and Mispec Plateau, both probably part of the Tertiary pediplain, suggest differential uplift and faulting between these areas. Similarly, the sharp termination of the pediplain at the Chaleur coast at elevations between 150 and 300 m suggests a tectonic origin for Chaleur Bay, although an erosional origin cannot be completely ruled out.

Other evidence of tectonism unrelated to postglacial warping is rare in the Cenozoic, although areas of high seismicity (Rast et al., 1979) are present (most notable being along the Oak Bay Fault) and postglacial reverse faults, attributed to thrusting from the southeast, have been reported at Saint John (Matthew, 1894a, b).

Present drainage patterns and valleys, which have developed penecontemporaneously and subsequent to uplift of the Tertiary pediplain, are in part structurally adjusted (i.e., their position determined by the location of the most easily eroded strata) and in part determined by local glacial history and nature and history of submergence and emergence.

Numerous examples of the development of structurally controlled subsequent valleys are present throughout New Brunswick. For example, most valleys in the Anagance Ridges and Shepody Ridges are controlled by streams eroding along belts of easily eroded strata and as such parallel the strike of major structural elements. Stream valleys on Jacquet Plateau parallel bedrock structure; their location is governed by bedrock competence. Broad basins and some major stream valleys in the Saint-Quentin Plateau are developed on the more easily eroded rocks of the Matapedia Group (Appendix 1). Curventon-Bathurst valley, which disrupts the gentle eastward slope of the Tertiary pediplain south of Bathurst, would appear to be a major valley that has developed along the strike of a Devonian/Carboniferous unconformity; stream erosion has been directed from the competent Ordovician strata along easily eroded lower Carboniferous shales and sandstones.

Parallel drainage patterns, developed in areas such as the Edmundston Highlands, St. Croix Highlands, and southern parts of the Miramichi Highlands, are largely due to the effects of glaciation as major valleys and ridges parallel glacial flow lines. Stream valleys on much of the Central Lowland are consequent in that they parallel the broad slope of the land and the dip of Pennsylvanian strata. They have only been diverted where glacial flow patterns did not parallel the general slope of the land.

Some major streams and valleys are transverse to major structural and physiographic elements. The most notable examples are the middle and lower course of Saint John River, the middle and upper parts of Nepisiguit River, and the lower course of Petitcodiac River. The general trend of these valleys has undoubtedly resulted from superimposition of ancient drainage courses upon structural and physiographic elements. Quaternary events may have affected parts of the

courses of these superimposed streams. Saint John River may have been diverted from an original course east of Grand Lake by the presence of late glacial ice on the Central Lowland during deglaciation. Similarily, the Petitcodiac may have been diverted from a course east of Moncton to its present position by late glacial ice persisting in the Northumberland Strait during deglaciation. The upper Nepisiguit system may well have formerly drained to the north via the gap at Upsalquitch Lake and been subsequently diverted to the east by late glacial ice persisting to the north of Upsalquitch Lake.

SURFICIAL MATERIALS

Surficial materials and landforms shown on Map 1594A are the result of past and present weathering, mass wastage, glaciation, marine and lacustrine submergence, isostatic rebound, paludification, alluviation, karst, stream and coastal erosion. Units have been defined that have a similar genetic origin, composition and thickness, and morphology. Areas of some materials or landforms that are too small to be mapped have been included in other map units.

Units are identified on the map by letters, which designate their compositional-genetic category and their morphology or thickness. The compositional-genetic categories are: **A** – Alluvial, **C** – Colluvial, **G** – Glaciofluvial, **L** – Lacustrine, **M** – Morainal, **O** – Organic, **R** – Bedrock, and **W** – Marine. Age differences between similar compositional-genetic units are shown by differences in colour hues on Map 1594A, the oldest unit generally having the darkest shade.

Suffixes to the compositional-genetic notations indicate morphology and thickness: **b** – blanket, generally 0.5 to 1.5 m thick, rarely thicker, morphology reflects relief of underlying bedrock surface; **m** – rolling (gentle to moderate slopes), rarely hummocky (numerous moderately steep to steep slopes present), generally greater than 1.5 m thick; **p** – plain, generally greater than 1.5 m thick; **r** – ridge (includes beach, bar, and spit), generally greater than 1 m thick; **x** – combination of landforms, hummocky to rolling but including planar surfaces, generally greater than 2 m thick; **v** – veneer, generally discontinuous and less than 0.5 m thick, morphology reflects relief of underlying bedrock surface. The thicknesses are subjective because they are based on a qualitative assessment of observed thicknesses of sediments in section and on the frequency of bedrock outcrop.

Textural prefixes, **a** – stony, **b** – bouldery, and **s** – sandy, have only been used to describe the morainal deposits, specifically tills. As few detailed textural studies of tills have been completed in New Brunswick, it was necessary to employ a classification that would characterize the morainal deposits qualitatively. Thus, the morainal deposits have been described according to their stone content and texture as depicted in Figure 3. Tills are loamy where no prefix is used to describe their texture. Silty and clayey tills are present, but occupy small areas that are not mappable at the present scale.

Pre-Quaternary

Bedrock (R)

The bedrock unit includes areas where rock is at or near the surface. Thin unconsolidated deposits, mainly glacial drift on upland surfaces, colluvium on slopes, and alluvium and organic deposits in low areas may be present. Extensive areas of surface or near-surface bedrock are also present in areas veneered by morainal deposits (Mv) or moraine deposits. Isolated occurrences of surface or near-surface bedrock occur in areas of thicker unconsolidated deposits, especially those mapped as morainal blanket (Mb), and marine blanket (Wb).

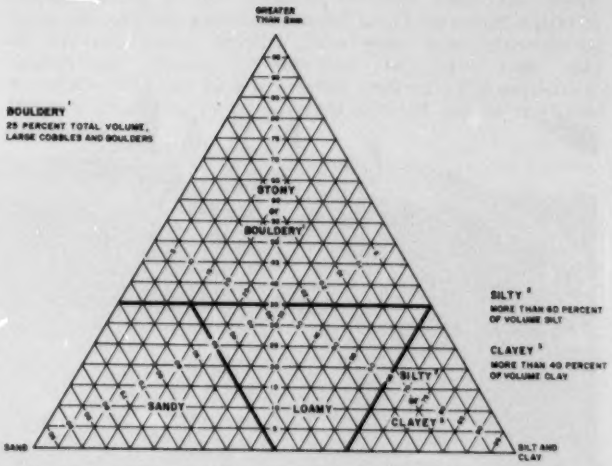


Figure 3. *Ternary diagram showing grain size composition of different textured tills.*

Distribution and Materials

Bedrock outcrops are mainly restricted to areas scoured by glaciers, to steep slopes where rock is exposed by mass wastage, and to areas stripped of unconsolidated materials by wave washing during submergence. Glacial scouring has produced abundant rock outcrops on ridges and hills with axes orientated transverse to glacial flow, for example Bocabec Hills, Clarendon Hills, Anagance Ridges. Extensive bedrock outcrop on the Jacquet Plateau, Campbellton Hills, and northeastern Miramichi Highlands is also probably due to glacial plucking and scouring. Mass movement has removed unconsolidated materials and weathering products on steep, stream- and wave-eroded slopes; such exposed bedrock is too small to map at the scale of Map 1594A. The latter are particularly common along the Fundy coastline of the Caledonian Highlands. Most streams in New Brunswick are entrenched and have rock walls. Glacial drift was apparently thin enough and wave washing intense enough on parts of the New Brunswick Lowlands that the drift has effectively been removed from some bedrock surfaces there (Rampton and Paradis, 1981b). Most hill and ridge tops along the southern edge of the St. Croix Highlands, bordering the Bay of Fundy, have been swept clean of unconsolidated materials by wave washing. Rock also forms the tors (Fig. 4) present on some of the higher peaks in the Miramichi Highlands (Gauthier, 1979, 1980b).

The bedrock surface is generally weathered where naturally exposed. Glacial polish is commonly seen only on freshly exposed bedrock surfaces. Grooves and streamlined landforms, such as whalebacks and roches moutonnées (Fig. 5), are preserved at some localities. Lineations seen on airphotos of some bedrock areas are the result of glacial fluting or of structure. Fragmented and weathered rock is common on bedrock surfaces and is difficult to distinguish from thin drift because clasts within thin drift are predominantly local, and are shattered and weathered in a manner similar to those directly dislodged from bedrock. The depth of weathering and disintegration is generally limited to the upper 0.5 m of rock (Rampton and Paradis, 1981a, b, c).

Bedrock overlain by unconsolidated materials commonly is unweathered. However, Carboniferous sandstones, which underlie most of the New Brunswick Lowlands, are decomposed and disintegrated in the subsurface to depths

Figure 4. *Tor on Colonels Mountain in Northern Miramichi Highlands; weathering pits are present on the tor.*

Figure 5. *Roches moutonnées along the west coast of Deer Island.*

between 0.5 and 2 m in many localities (Rampton and Paradis, 1981b; Gauthier, 1982). Typically "The lower half of the sequence is composed of broken-up sandstone clasts, loosely associated and parallel to the bedrock stratifications. No relative displacement of fragments is observed and the density of fracture increases upward. Fragments are sub-angular due to the weak cementation of the sandstone grains. In the upper half, fragmented layers of rock get thinner, with the presence of sand and silt beds between rock layers. In the upper part, sand becomes the dominant component and the top sequence grades toward a basal till, with a transitional contact. Compaction gradually increases, and rare erractics start to appear in the sand." (Gauthier, 1983). In some places the upper part of this sequence has been truncated or disturbed by glacial erosion (Gauthier, 1983).

Deep weathering and disintegration of bedrock have been reported at numerous localities throughout New Brunswick since the time of Chalmers (1887a, b, 1888, 1890). Granites weathered to depths of 2.5 m and Carboniferous conglomerates weathered to depths of 9 m have been noted by Gadd (1973) in southern New Brunswick. Disintegrated volcanics to depths of more than 8 m have been reported in the Caledonian Highlands (Rampton and Paradis, 1981c). Deeply weathered and altered granites have been noted throughout the Miramichi Highlands (Lee, 1957; Finamore, 1981; Anderson, 1968), especially in the Northern Miramichi Highland (Gauthier, 1979, 1980b, 1983) where Veillette and Nixon (1982) have drilled to a depth of 26 m in chemically altered granites and metasediments.

Age and Origin

The lithology, age, and origin of the various rock types exposed are given on numerous maps and reports describing the bedrock of New Brunswick (cf. Poole et al., 1970; Carroll, 1977; Potter et al., 1979; Fyffe, 1982a; Ruitenberg and Fyffe, 1982; Ruitenberg and McCutcheon, 1982). These descriptions are summarized in Appendix 1.

Bedrock surfaces showing glacial striations, grooving, and other erosional features generally have a Late Wisconsinan age. The age of weathered bedrock surfaces and the associated regolith, however, is more contentious. Chalmers (1887a, 1888, 1898) believed that disintegration of the Pennsylvanian sandstones predated Quaternary glaciation and that the regolith owed its preservation to ineffective erosion by glaciers. Gauthier (1983) also believed that the general sequence shown by the sandstones, whereby fractured bedrock grades upwards into sand, and sand grades upward into compact till, supports a preglacial origin. Rampton and Paradis (1981b) have raised the possibility that some of the sandstone fracturing may be the result of glaciation. The possibility also exists that some of the weathering may have been effected by periglacial events immediately after deglaciation and by groundwater and frost activity during the Holocene. Gauthier (1983) suggested that rock with high sulphide contents may be most susceptible to weathering during the Holocene.

The deep regolith developed on some granites may have developed over a great length of time, although Anderson's (1968) contention that weathering may have been initiated in the early Carboniferous may be extreme. Lee (1957, 1962a) and Gadd (1973) suggested that weathering of deeply weathered granites was initiated before the last glaciation because the weathered granite underlies till. Gauthier (1980b) suggested that the deep weathering on the granites of the Northern Miramichi Highlands indicated pre-Wisconsinan weathering. Wang et al. (1981) believed that soil characteristics of the highly weathered granite in this area also pointed to a pre-Wisconsinan origin, possibly pre-Quaternary. Seaman (1982b) noted the widespread occurrence of weathered rock and absence of pre-Wisconsinan tills, and suggested that much weathering of the granites may well have been initiated prior to Wisconsinan time, but probably is continuing today.

Wisconsinan and/or Pre-Wisconsinan

Colluvium and glacial deposits (M.Cb)

Colluvium and glacial deposits on high ground in the Northern Miramichi Highlands and Caledonian Highlands have been mapped as being possibly pre-Wisconsinan in age (Map 1594A) because of pedological and geomorphic characteristics of the deposits covering these two areas. Some question (as will be discussed later), however, exists as to whether these deposits have ever been affected by Wisconsinan glaciers or whether they have been only minimally affected by Wisconsinan glaciers. The deposits in the two areas may be of different age and will be discussed separately.

Figure 6. *Colluvium on slope southwest of Popple Depot. Weathered bedrock and material derived from slopewash and downslope creep are present.*

Materials and Origin

Northern Miramichi Highlands. Sandy and stony tills, colluvium, regolith, and weathered bedrock in the Northern Miramichi Highlands have been mapped as undifferentiated morainal and colluvial deposits (M.Cb and aM.Cb). A complex mixture of sandy till, colluvium, and disintegrated bedrock is present on flat and relatively gently sloping slopes, mainly on and along ridge and hill crests. Glaciofluvial deposits are likely widespread, but have been identified in only a few localities partly because of the similarity of texture of the till, regolith, and glaciofluvial materials, and partly because of the general inaccessibility of this area. Sand and gravel terraces along some valleys and meltwater channels are undoubtedly outwash, and a few knolls and areas of hummocky drift shown on Map 1594A consist of ice-contact gravels (Gx).

Colluvium, consisting primarily of weathered bedrock and reworked till, mantles slopes along the large, deeply

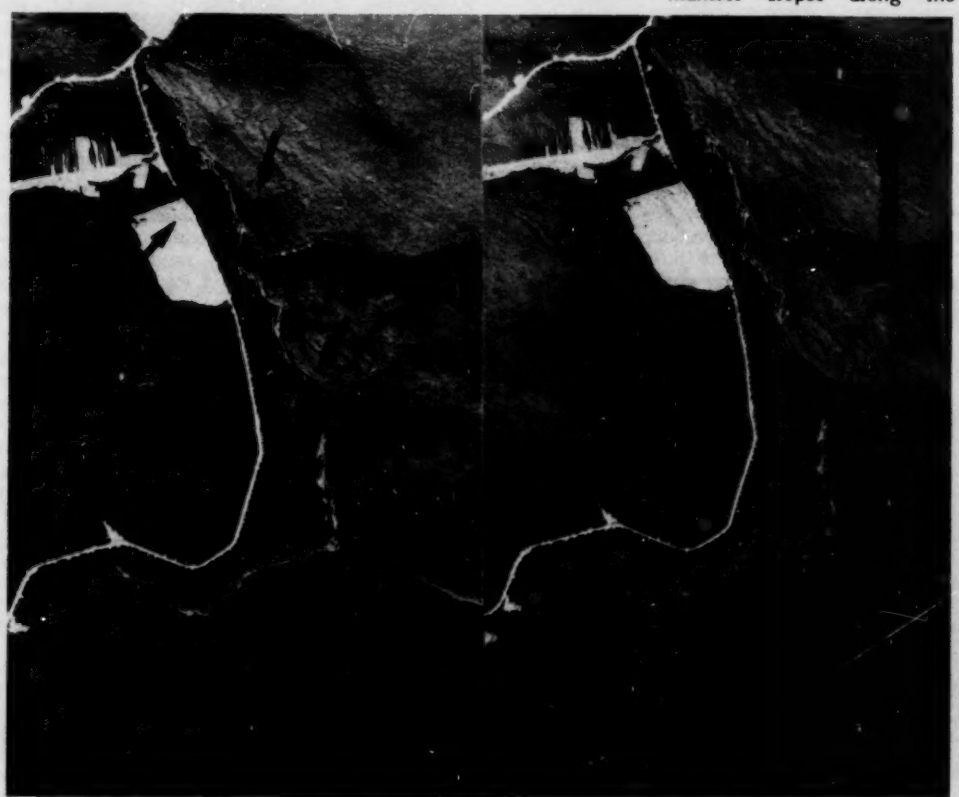


Figure 7. *Nivation terraces on slopes along West Branch of South Nepisiguit River. Arrows indicate nivation terraces (NAPL A25375-23, 24).*

entrenched valleys. Interbedded sandy gravels and diamictons consisting primarily of weathered bedrock (Fig. 6) indicate that active slope wash, creep, and possibly solifluction have occurred on the slopes. Gravel with clasts imbricated in an upslope direction overlying drift on benches, which have been tentatively identified as nivation terraces (Fig. 7); vertically oriented clasts; sequences showing a surface concentation of cobbles and boulders; and relict sorted polygonal ground (Fig. 8) all suggest that much of this unit has been affected by periglacial processes. Similarily, the intense oxidation of some soil profiles and their clay mineralogy (Wang et al., 1981) indicate that the unit has been subjected to a long history of subaerial weathering under climates either similar to today's or slighly warmer.

Generally, stony morainal and colluvial deposits (aM.Cb) cover areas underlain by Tetagouche Group metamorphics, whereas morainal and colluvial deposits with a large proportion of sand- and gravel-sized clasts (M.Cb) cover areas underlain by granitoid rocks. Surface concentrations of boulders have been noted in the sandy, gravelly unit. Unconsolidated deposits are generally more than 1.5 m thick throughout this area (Gauthier, 1979), although locally bedrock may be exposed. Areas of colluvium similar to those identified along the northwestern flank of the Central Plateau of the Caledonian Highlands (Cb on Map 1594A) may exist in the Northern Miramichi Highlands, but have not been identified or mapped.

Grus mantles much of the topography underlain by granites. Gauthier (1980b, p. 278) has described the grus as consisting "mainly of granule and sand-sized angular quartz and feldspar crystals. Mineral grains in the grus remain of similar size to those in the sound rock. Limited mechanical breakdown has affected the individual crystalline fragments; polycrystalline agglomerate of various minerals account for less than 5% of the total fragments in the 0.5 to 1.0 mm grain size fraction. On average, the grus consists of more than 75% sand and less than 10% clay. From mineral counts on 35 samples of grus (fraction 0.5 to 1.0 mm) and 5 thin sections of sound rock, the ratios of quartz to potassium feldspar to plagioclase were 4:4:1 for grus and 2.5:2.5:1 for sound rock. In the small grain size fractions, quartz concentration decreased as the plagioclase increased; this reflects the mechanical and chemical resistance of quartz, concentrated in the larger fraction (1 mm), as opposed to the chemical susceptibility of plagioclase." Veillette and Nixon (1982) have determined that the grus is at least 26 m thick in places.

Gauthier (1980b, p. 279) has identified a till unit within the grus based on pebble content, texture, and fabric: "Basal till was observed in most holes dug for seismic profiling. Systematic sampling of till and of underlying grus indicates that a high percentage of the erratic granules and pebbles in the till are fragments of parasedimentary, volcanic, and schist rocks, lithologies outcropping adjacent to the granite batholith. From preliminary results, a faint textural difference can be established between till and grus samples; granulometric curves show a lower sand content with respect to the silt and clay in the till, in contrast with a higher sand content in the grus. Ten sites were selected for till fabric studies which showed an "a" axis orientation (070-110°, dip 05-14°) in accordance with other regional ice flow indicators." The fabrics measured by Gauthier (1980b) would favour a glacial origin for the material identified as till even though some of the erratic granules and pebbles may be weathered out inclusions from within the granite as suggested by Veillette and Nixon (1982). The diversity of lithologies of boulders on Colonels Mountain and Big Bald Mountain also suggests glacial transport.

Caledonian Highlands. Undifferentiated stony morainal and colluvial deposits (aM.Cb) cover a large part of the Caledonian Highlands. The morainal component has been described by Rampton and Paradis (1981c, p. 7): "...as a blanket of stoney [sic] lodgment till (aMb). The till varies from sandy to rubbly to stoney [sic] to bouldery and areas of ablation till may also be present, but the different phases of till could not be subdivided at the scale of this investigation. In addition it is difficult to distinguish weathered and shattered bedrock from weathered till in some areas. The till thickness varies from negligible in areas of weathered bedrock to greater than 1.5 m in other areas . . . . most frequently the Central Plateau is covered by till with an average thickness of 0.5 to 1.5 m". Large areas of colluvium and rock outcrop are present along deeply entrenched valleys draining the Central Plateau (Rampton and Paradis, 1981c).

Very sandy material over shattered sandstone (sM.Cb) is present across a large area along the northern edge of the Central Plateau. The material may be sandy regolith over fragmented sandstone as suggested by Rampton and Paradis (1981c) or a very sandy till – a few exotic clasts are present within the material.

Small areas of ribbed moraine (aMm) are present; noteworthy is the McFadden Lake morainal belt, which is characterized by low parallel ridges (Rampton and Paradis, 1981c). Ice-contact gravels (Gx) and outwash have also been identified within the Central Plateau of the Caledonian Highland; most significant are the deposits at "Walton Lake and the head of the Big Salmon River drainage where a complex of kettled gravelly outwash, and esker ridges are present along a major meltwater channel crossing the central plateau" (Rampton and Paradis, 1981c, p. 8).

Figure 8. *Wedges of cobbles and boulders outlining sorted polygons in sandy till in Northern Miramichi Highlands.*

Age

The age of the colluvial and glacial deposits described above is problematic in both the Northern Miramichi Highlands and Caledonian Highlands. The two areas will be discussed separately as the unconsolidated deposits in them may have been deposited during different time intervals.

Northern Miramichi Highlands. Deposits mapped as Wisconsinan and/or pre-Wisconsinan in the Northern Miramichi Highlands cover a landscape having a definite nonglacial character, i.e., glacial landforms such as fluted bedrock, glacierized hillocks, and meltwater channels are absent. Valleys are deeply entrenched. Colluvium on the valley walls consists of interbedded slope wash and colluvial debris (Fig. 6), a characteristic not generally recognized in other areas of New Brunswick. Some antiquity for or relatively ineffective glaciation of the landscape is indicated by the above and the following: Tors are present on various mountains, most notably Big Bald, Colonels, and Sagamook (Gauthier, 1979, 1980b); soil development and rock weathering seem more intense than in any other part of New Brunswick (cf. Gauthier, 1980b, Veillette and Nixon, 1982; Wang et al., 1981); and periglacial features such as sorted polygons (Fig. 8) and nivation terraces (Fig. 7) are well developed. Periglacial features, particularly nivation landforms, suggest a long period of subaerial exposure of the ground surface under frigid conditions such as might be expected during the Wisconsinan. The highly oxidized nature of soil profiles, which requires a long time interval for its formation (Wang et al., 1981), suggests a pre-Late Wisconsinan age for the landscape. Alternatively, the oxidized and altered nature of the soils could be a function of the parent material.

Gauthier (1979, 1980b, 1983) suggested that this area was completely covered by cold-based ice during an early phase of Late Wisconsinan glaciation and that glacial erosion was ineffective, allowing phenomena resulting from earlier intense weathering to be preserved. The periglacial features are present, however, through a wide range of elevations and would require the presence of cold-based and noneroding ice through this range of elevations. The occurrence of glacially scoured terrain at similar elevations in areas immediately adjacent to the Northern Miramichi Highlands suggests that this is physically impossible.

Caledonian Highlands. Rampton and Paradis (1981c) have suggested an Early or Middle Wisconsinan age for the pre-Wisconsinan and Wisconsinan deposits covering the Caledonian Highlands. These deposits were assigned to the Wisconsinan because the striation patterns on the Caledonian Highlands have a similar trend to the patterns on a nearby marine rock bench assigned a Sangamonian age. The Early or Middle Wisconsinan age was assigned because: "The weathered and disintegrated nature of surfaces glaciated during the Caledonia phase relative to those glaciated during subsequent glacial phases; and the stream incision and the stream pattern development around the edge of the Caledonian Highlands all suggest relative antiquity.....Deeply disintegrated bedrock is occasionally present on the surface of the Caledonian Highlands. An extreme example is present near Londonderry where Precambrian volcanics (Ruitenberg et al., 1979) are broken and disintegrated to depths exceeding 8 m. Similar rocks to the southwest in areas glaciated subsequent to the Caledonia phase appear less shattered and broken, but this may be simply a function of rock lithology and jointing. The weathered, disintegrated and colluviated sandstone mantling Pleasant Mountain northwest of Elgin also appears relatively thick in comparison to similar sandstones in areas to the north glaciated by subsequent glacial phases. However, occurrences of deeply disintegrated and broken sandstone have been reported from the lowlands to the north (Chalmers, 1895). The periphery of the Caledonian Highlands glaciated only by the Caledonia phase and not by later glacial phases is characterized by deep stream incision and mature dendritic drainage patterns....Although this is in part a function of the high relief around the edges of the uplands, part of it may be due to the amount of time since the Caledonia phase. Adjacent areas with similar relief that have been covered by glaciers after the Caledonia phase do not show the same bold stream incision nor the well-developed dendritic pattern of drainage in the headwaters of streams". (Rampton and Paradis, 1981c, p. 19). Intensely oxidized soils have not generally been recorded for the Caledonian Highlands, but glaciofluvial deposits near Wolfe Lake on the north edge of Fundy National Park show intense oxidation and sorted polygons.

Wisconsinan

Units mapped as Wisconsinan consist of glaciofluvial and morainal deposits. Glaciofluvial deposits have been separated into those primarily deposited over or against ice (ice-contact deposits) and those deposited in a proglacial position (outwash). Morainal deposits have been separated according to whether they are primarily composed of lodgment or ablation till. Deposits and landforms whose age may predate the Late Wisconsinan are described together with known Late Wisconsinan deposits and landforms because of the lack of criteria to separate consistently the features of different ages.

Morainal Deposits

Morainal deposits of varying thickness cover much of the province. The term morainal deposits, rather than till, is an apt description for these materials because, even though they consist dominantly of till, in many places they comprise a complex of ablation till, lodgment till, glaciofluvial deposits, glaciolacustrine deposits, colluvium, and weathered bedrock (Rampton and Paradis, 1981a).

The morainal deposits have been separated on Map 1594A into those morainal deposits consisting primarily of lodgment till (Mb, Mv) and those consisting primarily of ablation till (Mm). The actual areal extent of each type of till may be more complex than that depicted on the map as the two tills are difficult to distinguish with certainty, especially where they are thin and may have been affected by soil-forming processes throughout their thickness (Rampton and Paradis, 1981a). Generally, the upper 0.5 m of any till is looser than the underlying till. This characteristic may be the result of thin ablation till being deposited over lodgment till or the result of postglacial subaerial processes such as soil-forming processes and slope erosion by running water (Gadd, 1973), slope wash, rill wash, and sheet wash during and subsequent to deglaciation (Lee, 1962a) or wave and current action in areas of submergence. Conversely, all compact till may not be lodgment till, but may locally be ablation tills of certain textures compacted by postglacial subaerial processes.

Lodgment Till (Mb, Mv)

Distribution and thickness. Morainal deposits consisting primarily of relatively compact till have been mapped as lodgment till. Generally, the surface of these deposits is flat, gently sloping, or reflects the topography of the underlying bedrock. Areas mapped as morainal veneer (Mv) generally have a discontinuous till mantle, less than 0.5 m thick, usually rich in local bedrock lithologies; bedrock outcrops may be common. Areas of thicker glacial drift and unconsolidated deposits may also be present, especially in areas of significant bedrock relief, such as in the Clarendon Hills (Thomas et al., in press); however, thickness rarely exceeds 2 m. Thin morainal deposits are most common in areas of high bedrock relief, which have probably been subjected to intensive glacial erosion, and areas well below the limit of marine submergence, which may have been stripped of much of their till cover by wave and current action.

Areas mapped as morainal blanket (Mb) generally have a till mantle greater than 0.5 m thick. Commonly till blanket is between 0.5 and 1.5 m thick (Rampton and Paradis, 1981a, b, c), but in some areas may be more than 1.5 m thick. Rampton and Paradis (1981a, b, c) have identified some areas in west-central and southeastern New Brunswick where tills appear to be generally thicker than 1.5 m. Gadd (1973) noted that till thicknesses are greatest in areas of drumlins; a drumlin field east of Port Elgin in eastern New Brunswick has also been mapped as thick till by Rampton and Paradis (1981b). Gauthier (1983) also noted greater than 12 m of till in fluted terrain south of Bathurst.

Texture. Textures and colours of tills are closely related to the underlying bedrock and bedrock immediately upglacier from the site of deposition. Tills in New Brunswick commonly have a loamy or a sandy loam texture reflecting the mixed nature of the bedrock having been overridden and eroded by the glaciers. Shales, sandstones, conglomerates, limestones, volcanics, igneous rocks, and their metamorphic equivalents all produce unique clast sizes when eroded and abraded by glaciers; loamy tills result where bedrock units are interlayered in a complex fashion as they are in New Brunswick. Shales, sandstones, limestones, and metasediments produce loamy tills with moderate stone contents, but igneous, volcanics and metasediments produce till with significant stone contents, generally exceeding 35%. Higher stone contents are even more prevalent where tills are relatively thin and where fractured weathered rock has been incorporated into the till through both glacial and postglacial processes.

Gauthier (1983) has characterized the tills of northern New Brunswick according to the texture of their matrix (Fig. 9). This figure indicates that the tills developed on igneous rocks and the Tetagouche Group are sandier loams than those developed on other sedimentary and metasedimentary rocks west of Curventon-Bathurst Valley.

Bouldery tills, or at least tills with boulder concentrations at the surface, are a common phenomenon in areas of conglomeratic and granitic rocks (Gadd, 1973; Rampton and Paradis, 1981a; Thomas et al., in press). In places where the boulders are concentrated only on the surface, they are probably an ablation component of the morainal deposits or may have become concentrated by removal of interstitial finer material by running water (Gadd, 1973) or by frost heave.

Although sandy loamy tills are common in New Brunswick, truly sandy tills only occur where glaciers have crossed sandstone strata with few beds of exposed shale, and where granite clasts, composing the till, disintegrate to sand-sized particles. The former are restricted mainly to areas underlain by Pennsylvanian sandstone (Rampton and Paradis, 1981b) and the latter to areas in the Northern Miramichi Highlands (Gauthier, 1980b).

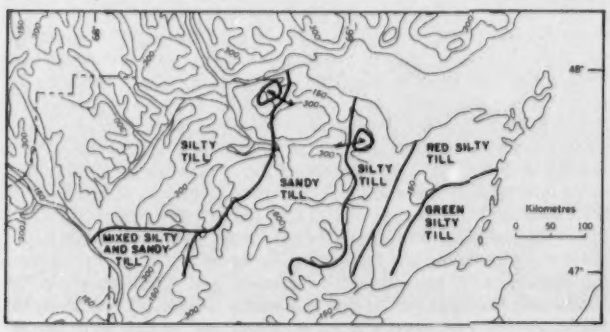


Figure 9. *Texture of matrix of tills in northern New Brunswick (modified from Gauthier, 1983).*

Clayey and silty tills are restricted to a few areas where poorly consolidated Carboniferous shales are exposed. Reddish clayey, silty tills have been reported from near Plaster Rock (Rampton and Paradis, 1981a), west of Fredericton (Lee, 1957), locally along Saint John River valley between Grand Falls and Keswick (Lee, 1959b, 1962a), and along the northeastern edge of the Acadian Peninsula (Gauthier, 1983) (Fig. 9).

Gauthier (1983) has made note of special textural variations within the tills present near Grand Falls. He reported pockets and discontinuous lenses of silt and crudely stratified sand within a predominantly silty (loamy) till. Lee (1955) mapped much of this area as rill wash (mixed sand and stones, reworked till) and pond deposits (clay, mixed clay, and stone). Ablation deposits and lodgment tills may be locally interstratified in this area.

A thin silty clayey diamicton has been noted overlying marine clays and underlying glaciofluvial deposits along the Chaleur coastline between Charlo and Jacquet rivers (Gauthier, 1983). This diamicton may be a till, but it is more likely a glaciomarine unit as stones scattered throughout the diamicton have no distinctive fabric and marine shells occur at its base.

Colour. Most surface tills in New Brunswick are oxidized to olive-brown, yellowish brown, or brown at the ground surface. Grey or olive-grey tills are more common along Saint John River valley, in the Chaleur Uplands, and in parts of southwestern New Brunswick (Lee, 1955, 1957; Gadd, 1973; Rampton and Paradis, 1981a, b, c; Gauthier, 1983). In parts of the New Brunswick Lowlands where green Pennsylvanian sandstone outcrops over extensive areas, the till has a greenish hue to it; Gauthier (1982) has mapped extensive tracts of greenish till on the Acadian Peninsula (Fig. 9). Locally, pinkish granitic and volcanic rocks give the tills a pinkish hue (Gadd, 1973).

Reddish brown to red tills are typical of areas with numerous outcrops of Carboniferous or Triassic sandstones and shales and of areas where glaciers advanced across reddish marine sediment. Reddish tills are found near Plaster Rock (Rampton and Paradis, 1981a), in the vicinity of and to the east of Bathurst (Gauthier, 1983), in areas scattered throughout the New Brunswick Lowlands where red shales outcrop (Rampton and Paradis, 1981b; Gadd, 1973), along the northern and eastern edges of the Caledonian Highlands, and near Sackville (Rampton and Paradis, 1981c). The red tills near Sackville were deposited by ice moving south out of Northumberland Strait where it incorporated red marine sediments.

Subsurface Tills. Multiple tills have been noted at a number of localities in New Brunswick. Lee (1957) reported interlayered red clayey till and buff sandy till near Fredericton. Gauthier (1978a) reported a grey till with a northeasterly trending till fabric overlain by a red till with a northerly trending till fabric in an exposure south of Bathurst. C.F. Gleeson (personal communication, 1982) reported the presence of multiple tills in mineral exploration trenches in the eastern St. Croix Highlands and Northern Miramichi Highlands. As reported by Lee (1962a), Finamore (1981), and Rampton and Paradis (1981a), multiple tills are a relatively common phenomenon along Saint John River and adjacent areas above the mouth of Nackawic River. Known localities are plotted on Figure 10.

Tills in the sections along the Saint John generally fall into two categories: "(1) those characterized by yellowish brown (e.g. 10YR5/4) colours and a relatively silty sandy

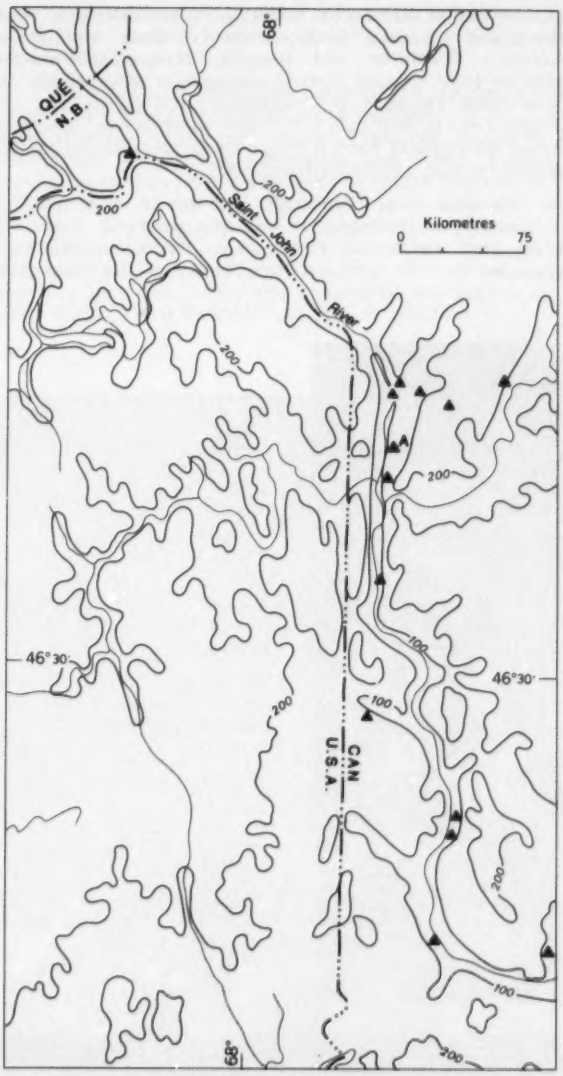


Figure 10. *Location of multiple till localities in upper and middle Saint John River valley (most localities from Lee, 1959b, 1961, 1962a; Kite, 1979; Rampton and Paradis, 1981a). See text for discussion.*

matrix and (2) those characterized by olive grey (e.g. 5Y4/2,3) colours and a relatively silty clayey matrix. The tills were difficult to characterize other than by colour and texture, although the grey till appeared to generally contain fewer limestones and calcareous pebbles, slightly more rounded and lenticular pebbles and more granitic exotics than the brown till. At sites where both tills were present, till fabrics in the grey till generally indicated an ice flow slightly north or east of the ice flow indicated by till fabrics in the brown till." (Rampton and Paradis, 1981a, p. 20).

Origin. As indicated previously, most deposits mapped as Mb and Mv are lodgment tills capped either by ablation drift or by till modified by soil development and other subaerial processes. Compact tills exposed in most sections are lodgment tills as they generally show a well defined fabric (Rampton and Paradis, 1981a).

Lee (1962a) suggested that some multiple tills are the result of different modes of glacier ice flow. Rampton and Paradis (1981a) suggested that the interfingering of grey and brown tills along Saint John River valley results either from two different sources of ice depositing till or from different modes of deposition occurring during one glaciation. It is difficult to attribute the grey and brown tills to two different sources as they both contain similar pebble types, a problem also encountered in northern Maine (Lowell, 1979). Till fabrics indicate that ice flow paralleled the river during deposition of both tills. An apparent higher content of granite clasts and more rounded pebbles in the grey till suggest that it had a more distant provenance, possibly a Laurentide origin. A similar provenance has been suggested for grey tills in adjacent Maine (Kite et al., 1982).

Age. Lodgment tills and associated glacial materials within this unit are, for the most part, assigned a Late Wisconsinan age. This is based on radiocarbon dates obtained on marine sediments interbedded with morainal deposits near Saint John and along the Chaleur coastline, radiocarbon dates on sediments overlying these tills, and the general deglaciation pattern (for further details see the section on *Quaternary History*).

In some areas flanking the cores of the Northern Miramichi Highlands and Central Plateau (Caledonian Highlands) and on the Acadian Penninsula, glacial deposits may predate the Late Wisconsinan. Glacial deposits covering the northern edge of the Central Plateau have been assigned a Wisconsinan age where the landscape has a smoother and more nonglacial aspect than in adjacent areas where glacial deposits are mapped as Late Wisconsinan. The morphological distinctions are less clear in similarly mapped areas to the south. Here deposits, mapped as Wisconsinan, cover areas characterized by a more complex history of sea level than areas covered by Late Wisconsian glaciers.

Glacial deposits assigned Late Wisconsinan and Wisconsinan ages in the area flanking the Bay of Fundy and on the Acadian Peninsula show no consistent differences in their postglacial modification or weathering. Glacial deposits assigned a Wisconsinan age near Nepisiguit Lakes are more oxidized than known Late Wisconsinan tills in adjacent areas (the oxidation might be simply a function of till lithological composition).

The above patterns suggests that tills adjacent to the core of the Northern Miramichi Highlands and the northern edge of the Central Plateau may well date from the Early or Middle Wisconsinan, whereas tills along the Bay of Fundy and on the Acadian Peninsula may be Late Wisconsinan, but predate the morainal complex at Saint John dated at ca. 13.2 ka. Stea (1983) has suggested a "Middle-Late Wisconsinan" age for equivalent tills in Nova Scotia adjacent to the Bay of Fundy.

Tills in the multiple till sections along Saint John River may range in age from Late Wisconsinan to pre-Wisconsinan. Undoubtedly the upper brown tills in all sections (whose upper sequence is not truncated) are Late Wisconsinan. Grey and brown tills underlying the upper brown tills may be Late Wisconsinan or may date from an earlier Wisconsinan interval. The latter is especially plausible where the upper brown till is separated from the lower tills by glaciofluvial beds. At one locality on Saint John River north of the mouth of Little River (site A, Fig. 10), a grey till, which is separated from a complex of tills higher in the section by 10 m of gravel, is oxidized in its upper 0.5 to 1 m (Rampton and Paradis, 1981a) and is thus possibly pre-Wisconsinan in age.

Ablation Till (Mm)

Ablation till usually has a similar texture to that of lodgment till, but has a slightly higher stone content and is less compact. Ablation till contains numerous pockets of ice-contact stratified material. Brinsmead (1977, p. 13) has described ablation till as consisting of "poorly sorted to largely unsorted material, usually containing, in significant percentages, component particles from the entire range of clasts sizes from clay through boulders. It is moderately to poorly compacted, the compaction factor normally being less than that of lodgment till, but may include pockets of better sorted ice-contact granular material".

Ablation till in areas underlain by granitic rocks may be stony and bouldery with extremely sandy and gravelly matrixes (Rampton and Paradis, 1981a; Gauthier, 1983), whether their original texture was such or whether the fines have been removed by meltwater activity is not clear (Thomas et al., in press). In these same areas, ice-contact stratified deposits form a major constituent of the morainal deposits (Gadd, 1973; Gauthier, 1983).

Ablation till generally exceeds 5 m in thickness; it varies in thickness from negligible to more than 10 m just north of Fredericton (Seaman, 1982b), and averages 10 m in northern New Brunswick (Gauthier, 1983).

Figure 11

Hummocky ablation moraine near Nigadoo Lake (NAPL A25727-49, 50).

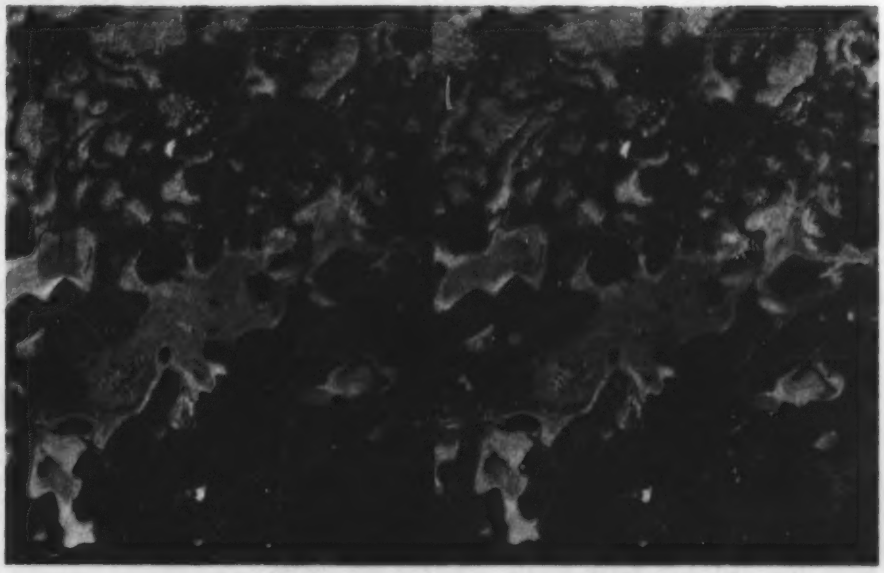


Figure 12

Sunbury moraine; hummocky deposits in depressions (NAPL A25368-36, 37).

Most investigations of granular resources in the province indicate that within small areas ablation till is generally thicker than lodgment till.

Topographically, moraines composed of ablation till are marked by rolling or hummocky topography (Fig. 11, 12); ribbed morainic topography, as described and illustrated by Prest (1968), is common. Rampton and Paradis (1981a, b) have illustrated numerous examples of ribbed moraine in the southern part of the Miramichi Highlands and the New Brunswick Lowlands; further examples from the St. Croix Highlands and northern part of the Miramichi Highlands are illustrated in Figures 13 and 14. The ribs or ridges are generally less than 15 m, and commonly less than 5 m, in height (Gadd, 1973; Rampton and Paradis, 1981b).

Hummocky, ribbed, and rolling moraine is generally present in areas where glaciers have formerly maintained a stationary or slowly retreating ice front or a zone of ablating dead ice. Hummocky, ribbed, and rolling moraine also appears to be concentrated in areas where granitic rocks have been glacially eroded and the resulting debris redeposited (for example, Central Miramichi Highlands, Pikiok Hills, Magaguadavic Highlands, and Clarendon Hills). Nevertheless, large moraines are also present in the New Brunswick Lowlands, most notably east of Oromocto Lake and near Gagetown (Gadd, 1973), near Sunbury (Rampton and Paradis, 1981a), at New Canaan west of Moncton (Brinsmead, 1974c), near Sackville (Rampton and Paradis, 1981c), along the axis of the Tormentine Peninsula (Rampton and Paradis, 1981b), and west of Tracadie on the Acadian Peninsula (Map 1594A). Ribbed moraine and rolling moraine are common in the valley bottoms of the Anagance

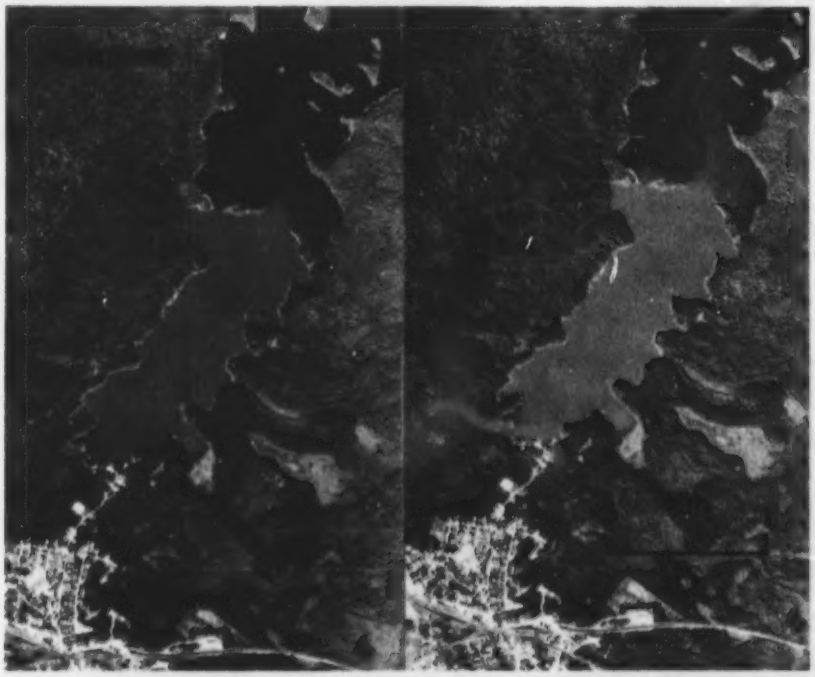


Figure 13

Ribbed morainic topography near McAdam (NAPL A23858-237, 238).

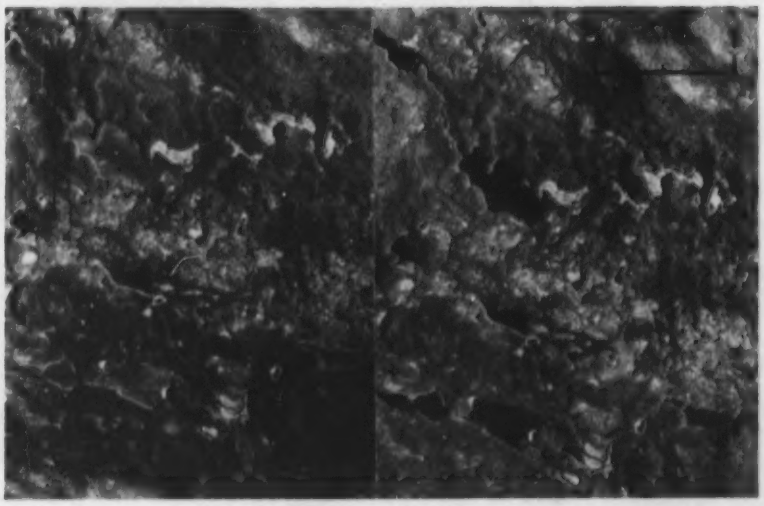


Figure 14

Ribbed morainic topography southeast of Serpentine Lake (NAPL A24434-23, 24).

Ridges, marking terminal positions of the retreating Late Wisconsinan glaciers. Small moraines are scattered throughout the province, most notable are those west of Plaster Rock (Rampton and Paradis, 1981a), west of Nigadoo (Finamore, 1978; Gauthier, 1978a), and south of Lorne (Brinsmead, 1979b; Gauthier, 1982).

DeGeer moraine or minor morainic ridges are common in the area between Newcastle and Bathurst. These ridges have minor relief, but are relatively short, straight, and regularly spaced (Fig. 15).

Origin. Moraines composed of ablation till result from downwasting of debris-rich glacier ice. End moraines and belts of thick ablation till would suggest a narrow zone of ablating ice being supplied with debris by active ice over a considerable period of time, in a manner similar to models proposed for deglaciation of much of New England (cf. Mulholland, 1982). In New Brunswick, especially in the Southern and Central Miramichi Highlands and in the western St. Croix Highlands, this model is supported by the fact that many ablation moraines are ribbed (or ridged) and that associated glaciofluvial systems commonly consist of an esker or series of eskers that terminate at the morainic belt. Outwash plains or valley trains extend downstream from the morainic belts.

Glacial ice may become isolated from active source areas when it thins in areas of high relief and thereby be inhibited or prevented from flowing across these areas. Thin layers of ablation till over broad areas may owe their existence to the gradual downwasting of dead ice, which has been isolated in this manner. Similarily patches of thick ablation moraine showing little sign of ridging or associated development of glaciofluvial systems may have been formed through this process. Seaman (1982b) has proposed such a model for much of the ablation drift near Fredericton to the east of the Miramichi Highlands. The numerous patches of ablation moraine and associated ice-contact deposits north of Grand Falls may also be the result of downwasting of an ice mass in its final phase of decay.

DeGeer moraine south of Bathurst on the Acadian Peninsula was probably formed by glaciers with ice fronts terminating in water bodies. The glacier margin, which was retreating in a southwest direction during deglaciation, would have ponded lakes between itself and bedrock to the north and/or glaciers to the east in high areas near the headwaters of Bartibog River north of Newcastle. The ice front to the west was most likely calving into a thin arm of the sea protruding up Curventon-Bathurst Valley and adjacent low ground.

Age. Most moraines composed of ablation till and related features are Late Wisconsinan in age. Major exceptions are the ablation moraines at Nepisiguit Lakes, west of Tracadie, and in a few of the valleys along the northwest edge of the Caledonian Highlands. As previously indicated, weathering and soil characteristics of glacial features at Nepisiguit Lakes suggest an age predating the Late Wisconsinan. Moraines west of Tracadie may predate the Late Wisconsinan or may simply belong to an early phase of it (Gauthier, 1983). Moraines in the Caledonian Highlands may vary in age from Late to Early Wisconsinan as previously inferred for the age of other glacial deposits in this same area.

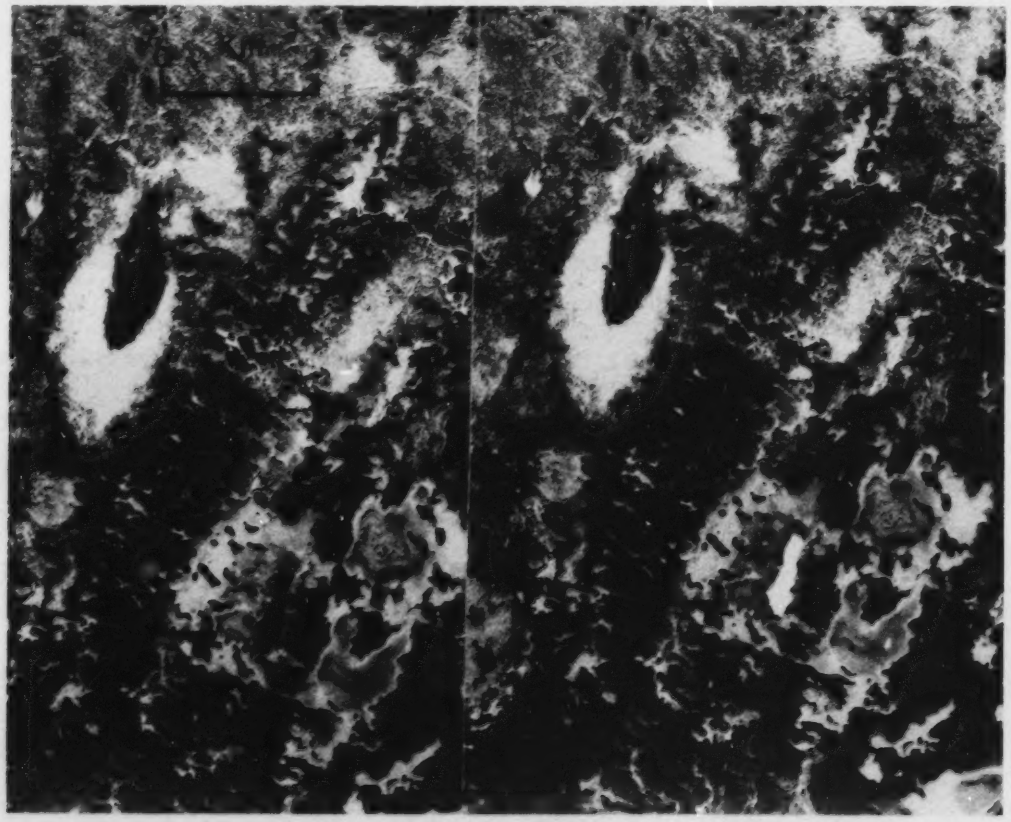


Figure 15. *DeGeer moraine near the headwaters of Bartibog River (NAPL A25375-11, 12).*

Figure 16. Large multiple-crested esker near Magaguadavic Lake (NAPL A23859-24, 25, 26).

Morainal ridges at Saint John seem to date ca. 13.2 ka and may mark the Late Wisconsinan limit. Glacial deposits along the Chaleur coast date ca. 12.4 ka. As final deglaciation of New Brunswick occurred by 11.5 ka, other moraines date in the interval between 11.5 and 13.2 ka. Further detail of ages of moraines is given in the section on *Quaternary History*.

Glaciofluvial Deposits

Glaciofluvial deposits are scattered throughout the province, but are locally concentrated. Glaciofluvial deposits have been divided into those deposits consisting of a complex of ice-contact deposits (generally dominant) and outwash (Gx), and those deposits consisting primarily of outwash (Gp). Ice-contact deposits are defined as those sediments deposited by meltwater on or against glacial ice and showing internal glaciotectonic deformation. Outwash has been limited to those sediments deposited by proglacial meltwater streams; generally they are devoid of glaciotectonic deformation. Glaciofluvil deltas (Gd) have been included within outwash deposits even though in many cases the proximal edge of the delta may have abutted against glacial ice during its formation. Detailed descriptions of many glaciofluvial deposits in New Brunswick have been completed during investigations of granular resources (see list of references in *Introduction*).

Ice-Contact Deposits (Gx)

Ice-contact stratified deposits have been described by (Brinsmead, 1977, p. 14) as characteristically comprising "well- to poorly-stratified sand and gravel that may include varying amounts of silt and clay. Ice-dropped boulders are common in this material. Stratification of ice-contact deposits is commonly contorted and faulted...Abrubt changes in grain size and degree of sorting are characteristic of such deposits as well." Gauthier (1983) and Rampton and Paradis (1981a) noted that ice-contact deposits commonly contain diamicton beds or ablation till, which is poorly sorted, but partially washed of fines. Large boulders have been noted as a common constituent of ice-contact deposits throughout New Brunswick (Gadd, 1973; Melvin, 1966). This common association led Gauthier (1982) to group many ice-contact deposits with ablation till in a unit designated as ridged and hummocky disintegration moraine.

In low areas adjacent to the coast, ice-contact deposits commonly contain large amounts of massive or stratified fine sand, silt, and clay showing evidence of glaciotectonic deformation. Melvin (1966) noted this association in areas covered by ice-contact deposits east of Saint John and Rothesay. Faulted and folded ice-contact complexes containing layers and lenses of fine sand, silt, and fine sand are also present west along the coast to Oak Bay near St. Stephen and south as far as Campobello Island. Ice-contact deposits along deeply entrenched valleys, such as the Saint John, Miramichi, and their tributary valleys, also contain ice-contact deposits characterized by large amounts of fine grained sediments or massive sands.

Ice-contact deposits are mainly present in the form of eskers, kames, kame terraces and deltas, and kame and kettle complexes. In some places these landforms are arranged in the form of end or interlobate moraines. In many places the above landforms, which are typically composed of stratified glaciofluvial deposits, have a muted nondescript surface expression. This generally occurs in areas that lie below the upper level of marine or lacustrine submergence, or along valleys where postdepositional processes have modified their original form.

Ice-contact deposits and landforms are generally concentrated in specific regions. Long segmented eskers, locally consisting of multiple-ridged crests, are most common in the Central and Southern Miramichi Highlands and western St. Croix Highlands, the most spectacular being the large esker east of Magaguadavic Lake (Fig. 16) and the system extending south of McDougall Lake (Gadd, 1973). Smaller eskers can be found in the Edmundston Hills, Jacquet Plateau, terrain within and adjacent to Curventon-Bathurst Valley, and southwestern Caledonian Highlands. A large esker occurs east of Blackville in the Central Lowland. Major kames and kame-and-kettle complexes can be found along Saint John River valley and its major tributaries west of the mouth of Keswick River, within valleys in the Anagance Ridges, west of Paquetville on the Acadian Peninsula, at 'Blue Mountain' southwest of Bathurst (Gauthier and Cormier, 1977), along the upper and middle reaches of Nepisiguit River valley, and within the Tobique Basin and adjacent Central Miramichi Highlands (Fig. 17). Two large, discontinuous interlobate kame moraines (Prest, 1968) border

Figure 17. *Eskers, meltwater channels, and ice-contact glaciofluvial deposits in Gulquac Mountains (NAPL A23858-159, 160).*

the Chaleur coastline west of Nepisiguit Bay (the one extending from near Balmoral to west of Durham Centre was termed the 'Restigouche Kame' by Chalmers (1887a)). Large morainal ridges consisting primarily of ice-contact stratified drift are present near Saint John (Gadd, 1973).

Extensive subtill glaciofluvial deposits, many of which may be of ice-content origin, occur west of Plaster Rock (Rampton and Paradis, 1981a) and along Saint John valley west of Grand Falls (Kite, 1978). These deposits are characteristically gravelly; those west of Plaster Rock contain Laurentide granites and gneisses.

Origin. Eskers in New Brunswick formed both in tunnels and channels within the glacier ice. Most eskers traverse the landscape with little regard for topography and undoubtedly formed in subglacial tunnels where meltwater flowed under a hydrostatic regime imposed by glacier ice. Indeed, Gadd (1973) has noted that segments of some major eskers define continuous glaciofluvial systems across major valleys such as the Saint John. Small narrow esker ridges that generally follow valley bottoms may have formed in either tunnels or narrow, open channels. A large ridge near Plaster Rock, which bifurcates into two tributaries (Rampton and Paradis, 1981a), probably was deposited in a channel developed in an interlobate suture.

Gadd (1973, p. 12) suggested the following sequence of esker development in the western St. Croix Highlands: "The western part of the area is characterized by a discontinuous system of ice-controlled drainages represented by eskers which, from the dips of strata in relatively few exposures and from the regional network pattern of the eskers, appear to have carried water southward across the easterly trending Saint John valley. Between the Saint John River and the Bay of Fundy, eskers trend southeasterly parallel to the regional trend of glacial flow. It is possible that the entire system was active at one time but the segmented nature of the system, the association with apparently successive major outwashes on the coast and along principal valleys, and the fact that esker ridges are found oriented more or less at right angles to moraine and minor moraine systems, where the relationship may be seen, all suggest that the abandonment of esker segments was sequential and controlled by topography and thus emphasizes the lobate configuration of the ice margin at different times. Some major eskers apparently were at some time continuous across major transcurrent valleys such as the Saint John; smaller eskers parallel to such valleys, and apparently younger than the major eskers, represent the transition from glacial to nonglacial drainage during the last local phase of ice-controlled glacial drainage."

Most isolated kames were deposited in isolated crevasses and voids in glacier ice, but some kame and kettle complexes may have originated as outwash plains and only took their form after melting of the glacier ice. Many kames, kame terraces, and kame deltas deposited within major valleys may have been deposited subaqueously as envisaged by Kiewiet de Jonge (1951) and Lougee (1954). Seaman (1982b) noted the occurrence of massive sand in association with stratified sand, coarse and cobbly sand and gravel beds, a few boulders, till-like diamicton, and faulted strata in a glaciofluvial deposit at Durham Bridge north of Fredericton; these characteristics generally indicate deposition in an ice-marginal subaqueous environment. Thick interbedded, folded, and faulted clays, silts, and sands east of Saint John and Rothesay, and along the Fundy coast west and south to Campobello Island were deposited subaqueously by meltwater on or against glacial ice as these deposits all lie well below the limit of marine submergence. Interestingly, marine sediments draped over these glaciofluvial sediments are faulted at a few localities. This indicates that buried ice blocks melted after general deglaciation of the area. Many sandy glaciofluvial deposits below marine limit along Miramichi River and its tributaries and on the New Brunswick Lowlands are considered to have a subaqueous origin also (Rampton and Paradis, 1981b).

The 'Restigouche Kame' and associated interlobate moraines along the Chaleur coast were deposited by meltwater in a suture formed between glacier ice on the Chaleur Upland to the south of the bay and glacier ice occupying Baie des Chaleurs. Rocks found within this system come from localities to the south and north of the kame (Gauthier, 1983). The general striation patterns confirm that separate, but coalescing ice sheets were present adjacent to the 'kame'. Those depositional features above levels of marine submergence were obviously formed in channels within the suture zone, whereas those at low elevation were formed subaqueously in channels flooded by marine waters, and subaerially as marine deltas.

The major end moraines in southern New Brunswick, which contain large amounts of glaciofluvial sediments, were formed mainly by meltwater streams debouching sediment into the sea at low elevations to form kames and kame deltas at distinctly defined ice margins. East of Saint John, where the ridge rises above sea level, it consists primarily of glaciofluvial sediments in the form of short esker segments and kames. Parts of the ridge appear to consist of ablation till either melted out of a debris-rich glacier ice or pushed into this position by this same ice.

Age. Subtill glaciofluvial deposits west of Plaster Rock and along Saint John valley predate the overlying till of Late Wisconsinan age; however, whether they were deposited during an early phase of the Late Wisconsinan or whether they predate the Late Wisconsinan is not clear. As indicated previously, subsurface tills in this area are also difficult to date.

Ice-contact deposits west of Paquetville on the Acadian Peninsula are Middle Wisconsinan or more likely date from an early phase of the Late Wisconsinan as do adjacent morainal deposits. Ice-contact deposits along the upper course of Nepisiguit River and areas along the flank of the Northern Miramichi Highlands may date from the earlier part of the Wisconsinan as indicated in the section discussing the age of morainal deposits.

Ice-contact deposits in the Saint John area and along the coast immediately to the west were deposited between 13.4 and 13 ka as active glacier ice was present at Saint John ca. 13.2 ka and much of the area was submerged ca. 13 ka. The 'Restigouche Kame' has been dated ca. 12.4 ka (see section on *Quaternary History*). Other ice-contact deposits date from between 13.4 and 11.5 ka according to their position relative to retreating ice margins during deglaciation (for more details see *Quaternary History*).

Outwash (Gp, Gd)

Brinsmead (1977, p. 15) has described outwash in general as being "typically well sorted and fairly well stratified gravel and sandy gravel. The content of fine particles is usually low in outwash deposits, as the meltwater from which the coarse material is deposited tends to carry the fines farther downstream"; he qualifies deltaic deposits as follows: "The base of a typical section comprises fine to silty fine sand with silt interbeds, grading upwards through medium sands and pebbly sand into medium-pebble to small-cobble gravel near the surface.... Stratification is much more regular and laterally continuous than that of outwash, which is more likely to exhibit cut-and-fill structures and truncated bedding." (Brinsmead, 1977, p. 16).

In the St. Croix Highlands, Gadd (1973) reported that coarse outwash forms the head of most valley trains and outwash plains. The materials of the major outwash deposits are similar to the ice-contact stratified drift and ablation material deposited at the ice front. Downstream the outwash is composed of "gravel with clasts of all sizes (including boulders), sand, and minor silt. Sorting of the material is fair to good and granules are mostly subangular to subrounded with roundness and abundance of resistant particles increasing rapidly from source" (Gadd, 1973, p. 14).

Outwash is mainly present in the form of outwash plains and fans, valley trains, and deltas. Locally these features may contain hummocks or hills where ice-contact deposits project through the flat to gently sloping outwash. Areas of ice-contact stratified materials deposited in subglacial or englacial meltwater channels commonly mark sites of meltwater flow and outwash deposition following the melting of glacier ice (see Frontispiece).

Extensive areas of outwash are present in the outwash plains and valley trains along Magaguadavic and Pocologan rivers in the St. Croix Highlands; as outwash fans northwest of St. Martins; and as valley trains on the upper reaches of Kennebecasis River in the Caledonian Highlands, along most major streams draining eastward and southward within the Miramichi Highlands, in the Campbellton Hills, along Green River north of Edmundston, and in meltwater channels on the Acadian Peninsula.

Deltas are common along the Bay of Fundy, especially between Lepreau and the Maine Boundary; along Baie des Chaleurs between the mouth of Upsalquitch River and Bathurst; near the mouths of major tributaries to Saint John River near Fredericton; and along Saint John River between Woodstock and Grand Falls. Major deltas are present at the mouth of Sevogle River and south of Little Southwest Miramichi River west of Newcastle. Other deltas are noted on Figure 18.

Subtill outwash has been mapped along Smith Creek south of Cornhill in the Aganance Ridges (Map 1594A), and has been noted in numerous exposures and boreholes along Saint John River north of Woodstock (Kite, 1979; Lee, 1961, 1962b; Rampton and Paradis, 1981a). Subtill gravels near Tracadie containing abundant clasts from the Miramichi Highlands (Gauthier, 1983) may also be outwash. (Subtill gravels and sands may, in some places, be alluvial (nonglacial) in origin.) Eastward dipping gravel and silty sand underlying a number of tills in a railroad cut southeast of Grand Falls (Rampton and Paradis, 1981a) may be of deltaic origin.

Origin. Most outwash plains and valley trains were deposited either within ice-walled valleys near the margin of glaciers or in depressions and valley bottoms immediately proglacial to the ice front. A classical example of glaciofluvial deposits being deposited by meltwater firstly on or within glacier ice, then in a proglacial valley probably ice-walled at its head, and finally in a deglaciated valley is present along Magaguadavic River between Magaguadavic Lake and Brockway (Map 1594A). In a few areas, topography caused outwash plains and fans to form a short distance from the actual ice front; for example, outwash fans at the base of the scarp marking the boundary between Mispec Plateau and Central Plateau west of St. Martins (Map 1594A) may have formed when the ice margin was a few kilometres to the northwest on the Central Plateau.

The outwash plain mapped at the head of Kennebecasis River may be in part deposited by a stream draining the Central Plateau and grading to a glacial ice dam in the lower part of Kennebecasis River (Rampton and Paradis, 1981c). Subtill outwash deposits in Smith Creek valley to the west may have formed in a similar manner.

Deltas are formed in a transitional environment where the fluvial environment impinges upon a marine, estuarine, or lacustrine environment. In New Brunswick, many deltas were deposited either in proglacial position with glacial ice forming their proximal boundary, or on stagnant ice resulting in isolated kettles on the delta surface. For example, many deltas along the Fundy coast west of Lepreau have ice-contact proximal slopes and kettles on their surface (Gadd, 1973). Other deltas, such as those along the Fundy coast between Saint John and Salmon River, near Keswick west of Fredericton, near the mouths of tributaries to the Saint John, and along Baie des Chaleurs (Fig. 18), formed a short distance from the ice front where swiftly flowing streams occupying braided channels encountered marine, estuarine, or lacustrine water bodies.

Large deltas seem to have formed only when an ample supply of sediment was being carried by glacial meltwater. A few small deltas in the Nashwaak River drainage system (Seaman, 1981, 1982b) and at the mouth of Mazerall Creek (Seaman and Thibault, 1981) may have formed subsequent to deglaciation. Other large areas of terrain, submerged to various elevations by marine, estuarine, and lacustrine waters following deglaciation, show no major deltas.

The large subtill deltaic sequence southeast of Grand Falls suggests deposition by meltwater from heavily debris-laden ice in a large lake to the east occupying the Tobique Basin; Tobique River was likely dammed this time.

Age. Subtill outwash in Smith Creek valley predates the overlying till of Late Wisconsinan age. Whether the outwash was deposited during an early phase of the Late Wisconsinan or predates the Late Wisconsinan is not clear. In some pits the gravels appear more oxidized than outwash of known Late Wisconsinan age and granitic clasts are weathered and disintegrating. These phenomena suggest a pre-Late Wisconsinan age.

Some outwash around the flanks of the Northern Miramichi Highlands and the north edge of the Central Plateau may be Early or Middle Wisconsinan in age. Ages have been assigned on their association with similarly dated morainal deposits. Outwash west of Paquetville on the Acadian Peninsula likely dates from an early phase of the Late Wisconsinan, but may also be Middle Wisconsinan in age.

Most outwash was deposited during the Late Wisconsinan after 13.4 ka. The large deltas west of Saint John slightly predate 13 ka, whereas outwash in the valleys to the north is slightly younger. Deltas associated with the 'Restigouche Kame' along Baie des Chaleurs and near Campbellton date from ca. 12.4 ka. Other outwash deposits date from between 13.4 and 11.5 ka according to their position relative to retreating ice margins during deglaciation (for more details see section on *Quaternary History*).

Late Wisconsinan and/or Early Holocene

Isostatic adjustment of land levels and eustatic changes in sea level caused much of New Brunswick to be covered by marine, brackish, or lacustrine waters during and following deglaciation. Gravel, sand, silt, and clay of varying thickness were deposited in these water bodies. A general paucity of fossils within these sediments has made it difficult to assign the sediments to a particular environment of deposition.

Deposits assigned a marine origin contain or are associated with sediments that contain marine fossils in adjacent areas. Assignment of deposits to this environment of deposition over much of the New Brunswick Lowlands, however, is tentative in that marine fossils are generally absent. Late Wisconsinan and early Holocene deglaciation patterns and relative sea levels indicate that some of these

nonfossiliferous sediments must have been deposited in the sea. Sand, silt, and clay deposited in Saint John River valley lack fossils. Late Wisconsinan and early Holocene deglaciation patterns suggest the presence of both marine and freshwater bodies in this valley. Thus most of these deposits have been assigned an undifferentiated lacustrine/marine environment of deposition.

Isolated deposits at levels well above the accepted maximum levels of marine submergence undoubtedly were deposited in glacially dammed lakes or lakes owing their existence to postglacial isostatic depression and tilting of the terrain.

Marine Deposits

Blankets and Plains (Wb)

Marine silty sand blankets much of the New Brunswick Lowlands adjacent to the Gulf of St. Lawrence (Map 1594A). In some places where glaciofluvial gravels have been reworked or on promontories exposed to the sea, more gravelly deposits, commonly in the form of bars and spits, have accumulated (Rampton and Paradis, 1981b). Early investigators (cf. Chalmers, 1895) referred to these deposits as the Saxicava sands. These deposits, which are here mapped as marine blanket, are generally 0.5 to 1.5 m thick, being thickest where glaciofluvial deposits were available for reworking. The marine blanket as mapped north of the Miramichi estuary may be thinner along its inland fringe than south of the estuary; all of Gauthier's (1982) nearshore sediments, which extend up to his marine limit, have been included in the marine blanket unit, whereas to the south deposits generally thinner than 0.5 m, which Rampton and Paradis (1981b) noted on submerged and wave-washed surfaces near marine limit, have been excluded from this unit.

Patches of clay and silt are present on the surface in broad flat areas. Clay and silt are interbedded with the sand and gravel on low areas near the present coastline. Clay and silt are commonly present in valleys and depressions where water well logs indicate significant thicknesses (Rampton and Paradis, 1981b).

Marine deposits are especially thick in the Bathurst Basin where up to 100 m of clay has been recorded in the subsurface (Gauthier, 1983). Generally the clays and silts in this area are covered by up to 2 m of sand and sandy gravel (Finamore, 1978). Significant thicknesses of marine deposits are also present along the Chaleur Coastal Plain (Gauthier, 1983). Here, glaciofluvial deposits – both ice-contact and outwash – have up to 2 m of marine sand and gravel on their crests, and greater thicknesses on their flanks. Valleys and low flat areas are commonly filled with more than 2 m of marine silt and clay. On gentle slopes facing Baie des Chaleurs these silts and clays are covered by littoral sands and beach gravels. Marine clays and a marine diamicton (or possibly till) have been noted under glaciofluvial deposits and younger marine deposits at a few localities along the coastline.

Thick marine deposits are present along the Bay of Fundy and east of Kennebecasis Bay just north of Saint John. Gadd (1973) noted that these deposits can generally be divided into a sand and gravel facies and a silty facies. The coarse textured facies is generally sand, up to 7 m thick, but bouldery gravel and boulder lag are also present. Gadd (1973, p. 15) described the silty facies and its relationship to the coarser facies as follows: "The marine silty sediments are most common in flat areas adjacent to the sea coast and at elevations commonly not more than 50 feet above present sea level. Thickness of up to 40 feet have been observed in actively eroding shore cliffs. Silt, silty clay, and clay in thin horizontal layers of one-half inch to several inches thick are common. In some of these there is colour banding with alternate dark grey to black bands; interbedding and interfingering relationships with sand and gravel deposits lying landward of the clayey deposits are exposed in a few localities. The most striking example of interfingering is exposed in road-cuts along service roads leading towards the modern beach in the large gravel pits at Sheldon Point. There, fossiliferous clay beds pinch out shoreward over a distance of about one thousand feet up the gradient of foreset beds of gravel and sand with which they are associated....

"Like the glacial tills of the area, the Pleistocene and modern marine clays have two principal colours, grey in the coastal areas west of Pennfield Station, and reddish brown in the areas eastward at least as far as the harbour of Saint John. These distinctive colours of similar sediment, apparently deposited simultaneously in the same marine basin, must reflect the provenance of the material and not differences in the marine environment of the Bay of Fundy."

Similar distributions of marine sediments are present east of Saint John along the Fundy coast, although they do not cover as much area (Finamore and Lohse, 1977). The coarser facies are primarily associated with glaciofluvial deltas. The finer facies are generally only present in the subsurface. Low benches consisting in part of thick sequences of marine gravels and sands overlying clay, silt, and fine sand containing pebbly diamictons and till beds are present east of St. Martins at Tufts Point (see Fig. 18 for location) and south of Martin Head. Multiple tills or till-like materials are also present within the lower, fine grained marine sequence at Red Head near Saint John.

Marine sediments underlie much of Saint John proper (Melvin, 1966) and terrain to the northeast of Kennebecasis Bay. In low areas silts and clays are generally capped by sand and gravel. East of Kennebecasis Bay sand and silt blanket much of the topography and form infillings in many depressions. Near Silver Falls just east of Saint John, the lower silts are convoluted in a fashion similar to the lower, fine grained marine sediments exposed in pits west of Mispec on the coast. Wedges of coarse sand and gravel cap many ridges and knolls.

Veneer (Wv)

Thin discontinuous marine deposits (generally less than 0.5 m thick), consisting mainly of silty sand, mantle much of the land surface below the limit of wave washing, but above the limit of a continuous recognizable marine blanket. Rampton and Paradis (1981b, p. 11-12) have described such areas as having "a discontinuous veneer of sand over the underlying material. The sand veneer may contain pebbly and silty layers. Pockets and lenses of silt and clay are present in some swales and environs protected from former intense wave-wash; and patches of sandy to pebbly gravel occur on hillsides and hilltops exposed to former intense wave-wash. Locally small areas of sediment thicker than 0.5 m are present."

Origin

Fine grained marine sediments indicate deposition in deep waters or in environments protected from high energy wave, tidal, or current action. Thus most marine clays and silts lie at low elevations, deposited when relative sea level was near its maximum elevation. Gadd (1973) noted that they were rarely found above 15 m elevation in southern New Brunswick. Clays and silts are common at a wider range of elevations in protected valleys; for example, thin clay and silt beds of supposed marine origin are present up to elevations of 60 m along Kennebecasis Bay, and 50 m along a valley north of Riverside-Albert (Fig. 18).

Table 2. Radiocarbon dates and descriptions for southern New Brunswick

Lab Dating Number	Date (BP)	Location	Elevation (m)	Comments	Reference
Pre-Late Wisconsinan					
BGS-806	>35 000	45°44.6'N, 65°32.2'W	40	Peat and organic silt from 1.2 m interlayered till and "greenish" sand underlying 3.5 m till and overlying 4.5 m weathered gravel and sand	
GSC-1680	>43 000	45°54'N, 64°39'W	10 (?)	Peat layer in silt sequence in sinkhole	Rampton and Paradis, 1981c Blake, in press
Late Wisconsinan Deglaciation					
GSC-3354	13 900 ± 620	45°13.5'N, 66°06.5'W	45	Marine shells (**Hiatella arctica**) from diamicton overlain by marine sediments and underlain by glaciofluvial sediments	
GSC-1222	13 600 ± 220	45°54'N, 64°39'W	10 (?)	Mastodon bone in silt sequence in sinkhole	Rampton and Paradis, 1981c Blake, in press
GSC-3557	13 100 ± 160	45°13.5'N, 66°06.5'W	45	Marine shells (**Hiatella arctica**) from diamicton overlain by gravel and sand	
GSC-2787	13 100 ± 280	45°35'N, 65°35'W	135	Basal silty gyttja from core 744-749 cm below lakewater/sediment interface	Rampton and Paradis, 1981c
GSC-882	13 000 ± 240	45°05.8'N, 66°45.3'W	40	Marine shells (**Portlandia** sp.) in sediment near base large delta	Gadd, 1973 Lowdon and Blake, 1970
GSC-2777	13 000 ± 330	44°38.9'N, 66°48.9'W	4 m above high water on spring tide	Shells (**Macoma calcarea**) from sand 30 cm over till at Red Point Beach, Grand Manan Island	Legget, 1980 Lowdon and Blake, 1980
BGS-684	12 700 ± 200	46°01.1'N, 67°32.4'W	105	Thin organic layer near top of stony clay under 1.3 m of sandy diamicton; original reported date of 14 750 ± 200 BP incorrect due to erroneous laboratory calculation	Rampton and Paradis, 1981a
GSC-1067	12 600 ± 270	45°15.3'N, 67°19.8'W	89	Gyttja from 639-644 cm below lakewater/sediment interface; lake level at 106 m	Lowdon and Blake, 1975 Mott, 1975b
GSC-2814	10 500 ± 170	46°9.5'N, 64°17.5'W	25	Basal organics from sediment core under Poulette Lake	Rampton and Paradis, 1981b
Late Wisconsinan and Early Holocene Submergence and Emergence					
GSC-2573	14 400 ± 330	45°13.5'N, 65°57.2'W	35	Valve of **Hiatella arctica** from 10 m below surface in sand pit	Rampton and Paradis, 1981c Lowdon and Blake, 1979
I(GSC)-7	13 325 ± 300	45°13.5'N, 66°06.5'W	3	Marine shells from clay interfingering upslope with gravel	Gadd, 1973 Lowdon and Blake, 1970 Walton et al., 1961
GSC-965	13 200 ± 200	45°13.5'N, 66°06.3'W	6 m above high tide	Marine shells from organic layers in clay interfingering upslope with gravel	Gadd, 1973 Lowdon and Blake, 1970
GSC-1340	13 000 ± 170	45°13.5'N, 66°06.3'W	9 m above high tide	Marine shells from clay interfingering with gravel	Gadd, 1973 Lowdon and Blake, 1970
GSC-3167	13 000 ± 200	45°13.7'N, 65°57.1'W	30	Marine shells (**Portlandia arctica siliqua** (Reeve)) from sand overlying clay and underlying 10 m of sand and gravel	Rampton and Paradis, 1981c
GSC-2640	12 800 ± 120	45°13.5'N, 65°57.2'W	30	Shell fragments of **Mya** sp. from 10 m below surface in sand pit	Lowdon and Blake, 1979 Rampton and Paradis, 1981a
GSC-3579	12 800 ± 140	45°13.7'N, 65°57.1'W	30	Shells from base of 3.5 m of contorted clay and silt underlying 10.5 m of sand and silt (contorted in bottom part)	
GSC-886	12 300 ± 160	45°12.6'N, 67°12'W	25 m above high tide	Marine shells (**Mytilus edulis** and **Macoma** spp.) from silt between glaciofluvial and beach gravel	Gadd, 1973 Lowdon and Blake, 1970

Table 2 (cont.)

Lab Dating Number	Date (BP)	Location	Elevation (m)	Comments	Reference
Late Wisconsinan and Early Holocene Submergence and Emergence (cont.)					
GSC-795	12 300 ± 160	45°08.5'N, 67°07.8'W	15 m above high tide	Marine shells in silt underlying glaciofluvial gravels and overlying offlap sediments	Gadd, 1973 Lowdon and Blake, 1968
BGS-666	9300 ± 140	45°57.2'N, 64°12.2'W	55	Basal 0.02 m of 1.6 m of peat overlying sand	Rampton and Paradis, 1981c
BGS-804	9050 ± 250	45°13.2'N, 64°14.5'W	45	Thin peat layer underlying about 0.4 m sand and 1.5 m of peat and overlying 10 m of sand and gravel	
GX-5363	8255 ± 350	45°12'N, 67°04'W	31	Gyttja from 515-518 cm below lake-water/sediment interface; lake sill at 49 m	Scott and Medioli, 1980
GSC-2628	7900 ± 120	45°10.2'N, 67°05'W	27	Basal gyttja about 345 cm below lake-water/sediment interface; lake level at 36 m	Lowdon and Blake, 1981
BGS-642	7880 ± 120	45°37.7'N, 64°50.5'W	18 m above high tide	Peat from base of bog overlying 17 m of glaciomarine sand and silt	Rampton and Paradis, 1981c
GX-5362	7535 ± 230	45°41.5'N, 64°43.5'W	-2	Gyttja from 210-212 cm below lake-water/sediment interface; lake level at 8-10 m	Scott and Medioli, 1980
GSC-3594	4500 ± 150	45°12'N, 66°22.5'W	41	Plant fragments from top of clay overlain by 1.8 m of peat	
GX-5364	6340 ± 340	45°10'N, 67°05'W	29	Gyttja from 604-606 cm below lake-water/sediment interface; lake sill at 36 m	Scott and Medioli, 1980
GX-5365	3120 ± 260	45°10'N, 67°00'W	-3	Gyttja from 306-308 cm below lake-water/sediment interface; lake sill at 8.4 m	Scott and Medioli, 1980
Late Holocene Submergence					
GSC-1089	4120 ± 130	45°50.8'N, 64°17.7'W	11.6 m below high tide	White pine stump rooted in till and salt marsh sediments	Lowdon and Blake, 1970 Rampton and Paradis, 1981c
GSC-930	4040 ± 130	45°51.1'N, 64°18.1'W	11.9 m below high tide	Stump rooted in till and overlain by salt marsh sediments	Lowdon and Blake, 1970 Rampton and Paradis, 1981c Grant, 1970
GSC-975	3520 ± 140	45°51.1'N, 64°18.1'W	9.4 m below high tide	Stump rooted in till and overlain by salt marsh sediments	Lowdon and Blake, 1970 Rampton and Paradis, 1981c Grant, 1970
I-3238	3430 ± 100	45°53'N, 64°21'W	7.3 m below high tide	Wood in salt marsh sediment sequence	MacNeill, 1969 Rampton and Paradis, 1981c
GSC-2718	3300 ± 60	44°37.2'N, 66°42.4'W	4 m below high water on spring tide	Tree root (**Larix laricina**) from submerged forest at Long Point Beach, White Head Island	Legget, 1980 Lowdon and Blake, 1980
I-3241	3290 ± 100	45°53'N, 64°21'W	7.3 m below high tide	Wood in salt marsh sediment sequence	MacNeill, 1969 Rampton and Paradis, 1981c
BGS-807	2600 ± 120	46°00.2'N, 64°33.3'W	4 m below high tide	Peaty layer from near base of 0.3 to 0.9 m thick clay layer with logs overlying 0.2 m sand and till and underlying peat and clay; peat contains no marine fauna	unpublished GSC Fossil Arthropod Report 81-16 by J.V. Matthews, Jr.
I-4313	1875 ± 95	45°57'N, 64°20'W	6.1 m below high tide	Wood stump from 6.1 m below surface of marsh sediment	Rampton and Paradis, 1981c Welsted, 1971

Table 2 (cont.)

Lab Dating Number	Date (BP)	Location	Elevation (m)	Comments	Reference
Late Holocene Submergence (cont.)					
GSC-1030	1760 ± 140	45°51'N, 64°18'W	6.1 m below high tide	Shells (**Mya arenaria**) 5.3 m below top of intertidal muds	Lowdon and Blake, 1970 Rampton and Paradis, 1981c Grant, 1970
BGS-667	1600 ± 90	46°00.2'N, 64°33.3'W	3 m below high tide	Peat from base of 0.3 m of peat overlain by 0.15 m of woody peat and 2.5(?) m of clay	Rampton and Paradis, 1981b
BGS-685	1460 ± 100	46°00.2'N, 64°33.3'W	3 m below high tide	Stumps under 2.5(?) m of clay; original reported date of 3500 ± 100 BP incorrect due to erroneous laboratory calculation	Rampton and Paradis, 1981b
I-4312	935 ± 95	45°57'N?, 64°15'W?	2.1 m below high tide	Organic material 2.1 m below surface of marsh sediments	Rampton and Paradis, 1981c Welsted, 1971
GSC-967	930 ± 130	46°05.4'N, 63°47.8'W	1-1.2 m below high tide	Stump under peat	Lowdon and Blake, 1970 Rampton and Paradis, 1981b Grant, 1970
GSC-602	640 ± 130	45°57'N, 64°20'W	0.3 m below high tide	Birch stump rooted in till and under one foot of salt marsh mud	Lowdon and Blake, 1970 Rampton and Paradis, 1981c
Submergence of Saint John River Valley (Marine, Estuarine, or Lacustrine)					
BGS-688	11 300 ± 200	46°01.1'N, 67°32.4'W	105	0.1 m of organics (macerated wood), 0.45 m from top of clay under sandy diamicton; original reported date of 13 350 ± 200 BP incorrect due to erroneous laboratory calculation	Rampton and Paradis, 1981a
BGS-663	10 180 ± 150	46°07.1'N, 67°16.2'W	135	0.05 m of peat underlying 0.5 m of clay and peat and overlying 0.5 m of clay and gravel on slope of ridge	Rampton and Paradis, 1981a
BGS-665	10 070 ± 150	46°01.1'N, 67°32.4'W	105	0.15 m peat lens from lenticular body of clay and peat between grey till (?) and roadbed	Rampton and Paradis, 1981a
Alluvial Events					
BGS-668	1725 ± 90	45°47.5'N, 64°30.2'W	15	Wood detritus in 0.5 m sandy unit overlain by 0.6 m of clayey silty sand and peat	Rampton and Paradis, 1981c
BGS-664	360 ± 80	45°59.2'N, 66°41'W	10	Wood from organic mat under 0.5 m of sand and 2.3 m of interbedded clay, silt, and sand	
BGS-669	modern	45°47.5'N, 64°30.2'W	15	Woody peat underlying 0.3 m of sand and overlying 0.7 m of sand and clayey silt	Rampton and Paradis, 1981c
Peat Accumulation					
BGS-805	6280 ± 170	45°18.5'N, 65°58.2'W	55	Layer from base of 1.8 m of peat overlying sand and gravel	
GSC-2524	6080 ± 80	45°37.2'N, 65°03.6'W	380	Peat from 175-185 cm below surface of bog	Lowdon and Blake, 1978 Rampton and Paradis, 1981c
GSC-2544	830 ± 70	45°37.2'N, 65°03.6'W	380	Peat from 87-98 cm below surface of bog	Lowdon and Blake, 1978 Rampton and Paradis, 1981c
Pollen Profiles					
GSC-1645	11 300 ± 180	45°15.3'N, 67°19.8'W	90	Gyttja from 621-626 cm below lake-water/sediment interface; lake level at 106 m	Lowdon and Blake, 1975 Mott, 1975b
GSC-1643	9460 ± 220	45°15.3'N, 67°19.8'W	91	Gyttja from 507-512 cm below lake-water/sediment interface; lake level at 106 m	Lowdon and Blake, 1975 Mott, 1975b

Table 2 (cont.)

Lab Dating Number	Date (BP)	Location	Elevation (m)	Comments	Reference
Pollen Profiles (cont.)					
GSC-3452	9320 ± 120	45°35'N, 65°35'W	135	Gyttja from 650-655 cm below lakewater/sediment interface	R.J. Mott, pers. comm., 1983
BGS-74	8684 ± 87	45°55'N, 66°40'W	120	Peat and organic detritus from depth of 610-620 cm in bog	Korpijaakko, 1976
BGS-77	8160 ± 76	46°02'N, 66°20'W	110	Peat detritus from 4.5-4.6 m below surface of bog near base of organics	Korpijaakko, 1976 Rampton and Paradis, 1981a
BGS-154	7920 ± 70	46°18'N, 65°08'W	90	Peat from 445-455 cm below surface of bog	Korpijaakko, 1976 Rampton and Paradis, 1981b
BGS-152	7860 ± 60	46°10'N?, 67°29'W	150	Organic detritus from 3.1-3.2 m below surface of bog near base of organics	Korpijaakko, 1976 Rampton and Paradis, 1981a
BGS-75	7222 ± 67	45°10'N, 66°38.5'W	85	**Sphagnum** peat from depth of 480-490 cm in bog	Korpijaakko, 1976
GSC-1095	7130 ± 140	45°09.2'N, 66°43.5'W	56	Basal gyttja from 1020-1025 cm below lakewater/sediment interface; lake level at 66 m, lake sill at 68 m	Gadd, 1973 Mott, 1975b
GSC-1513	6530 ± 210	45°15.3'N, 67°19.8'W	93	Gyttja from 347-352 cm below lakewater/sediment interface; lake level at 106 m	Lowdon and Blake, 1975 Mott, 1975b
GSC-1595	5120 ± 220	45°15.3'N, 67°19.8'W	94	Gyttja from 260-265 cm below lakewater/sediment interface; lake level at 106 m	Lowdon and Blake, 1975 Mott, 1975b
GSC-1693	3020 ± 150	45°15.3'N, 67°19.8'W	95	Gyttja from 160-167 cm below lakewater/sediment interface; lake level at 106 m	Lowdon and Blake, 1975 Mott, 1975b
Apparently Invalid Dates					
GSC-1063	16 500 ± 370	45°08.6'N, 66°42.8'W	52	Marl and gyttja from 622-629 cm below lakewater/sediment interface; lake level at 64 m	Gadd, 1973 Lowdon and Blake, 1975 Mott, 1975b
GX-6801	15 820 ± 770	45°14.5'N, 67°02.5'W	55	Organic mud from 448-451 cm below lakewater/sediment interface; lake sill at 65-70 m	Scott and Medioli, 1980
GX-6808	15 210 ± 1000	45°41.5'N, 64°43.5'W	-4	Organic mud from 390-410 cm below lakewater/sediment interface; lake level at 8-10 m	Scott and Medioli, 1980
GX-6805	15 125 ± 525	45°12'N, 67°04'W	30.5	Organic silt from 550-570 cm below lakewater/sediment interface; lake sill at 49 m	Scott and Medioli, 1980
GSC-1272	14 300 ± 270	45°08.6'N, 66°42.8'W	53	Gyttja from 588-593 cm below lakewater/sediment interface; lake level at 64 m	Gadd, 1973 Lowdon and Blake, 1975 Mott, 1975b
GX-6804	14 085 ± 260	45°10'N, 67°00'W	-4	Organic mud from 380-400 cm below lakewater/sediment interface; lake sill at 8.4 m	Scott and Medioli, 1980
GX-6053	12 385 ± 375	45°10'N, 67°05'W	28.5	Organic silt from 720-730 cm below lakewater/sediment interface; lake sill at 36 m	Scott and Medioli, 1980
GSC-1447	9080 ± 200	45°08.6'N, 66°42.8'W	55	Gyttja from 392-397 cm below lakewater-sediment interface; lake level at 64 m	Lowdon and Blake, 1975 Mott, 1975b
GSC-1604	6440 ± 300	45°08.6'N, 66°42.8'W	57	Gyttja from 230-235 cm below lakewater/sediment interface; lake level at 64 m	Lowdon and Blake, 1975 Mott, 1975b

Coarse grained sediments generally are the result of wave action in a high energy environment, especially where the waves are attacking glaciofluvial gravel and sand. Beaches and wedges of gravel are commonly developed on the crests of kames and glaciofluvial deltas. Wave and current action generally distributes sandy material over a broad area from the glaciofluvial deposits, as noted around the Blue Mountain ice-contact complex just southwest of Bathurst (Gauthier, 1983) and on the New Brunswick Lowlands (Rampton and Paradis, 1981b).

Some marine sediments were deposited in a glaciomarine environment as they are interlayered with till and till-like diamictons (e.g., near Saint John and along the Fundy coast) or interfinger with gravels and sands at the base of large deltas (Gadd, 1973). Stony diamictons overlying silt and clay along Baie des Chaleurs suggest that there, the marine sediments were deposited immediately in front of an advancing ice mass or under its floating margin. Numerous occurrences of clayey silt and fine silty sand within ice-contact deposits along Bay of Fundy and Baie des Chaleurs indicate that some marine sediments were deposited within and along the glacier front in association with glaciofluvial materials. Locally marine sediments were draped over glacial materials still containing buried ice blocks as shown by faulted marine sediments overlying ice-contact deposits near St. George.

Some sediments mapped as marine along Kennebecasis Bay and other valleys of the Anagance Ridges could be glaciolacustrine as glaciers may have blocked the mouths of these valleys when their middle segments were free of ice (Melvin, 1966; Finamore and Lohse, 1977; Rampton and Paradis, 1981b, c).

The sand wash over the New Brunswick Lowlands has been mapped as marine blanket and marine veneer in spite of paucity of marine fossils. Rampton and Paradis (1981b, p. 12) have attributed this phenomenon to the following "(a) the proximity of waning glaciers supplying large quantities of relatively cold fresh turbid water, (b) the relative shallowness of the waters covering part of the submerged area (the eastward slope of today's land surface is very similar to the slope of the postglacially upwarped surface of maximum submergence), (c) a land-sea configuration consisting of long estuaries with relatively fresh water along much of their length, and (d) the probable destruction of calcareous fossils because of the general absence of lime to act as a buffering agent, and the permeability of the porous sands and sandstones over most of the Maritime Lowland (Chalmers, 1895, 1902)." Some of the sand and associated sediment may have a glaciolacustrine or lacustrine origin as explained in the section discussing *Quaternary History*.

Age

Radiocarbon dates on bones and shells in sediments along the coastlines indicate that most marine sediments date from between 13.4 and 12 ka (Fig. 19; see *Quaternary History*). Some marine sediments on the New Brunswick Lowlands may be as old as 13.4 ka as the area likely was deglaciated between 13.4 and 13 ka (Fig. 20).

Marine sediments underlying fossiliferous till dated at 13 100 ± 160 BP (GSC-3557, Table 2) at Sheldon Point near Saint John, and underlying the lower till at Red Head just southeast of Saint John may be slightly older than 13.4 ka. Marine sediments, interbedded with and underlying till units at Tufts Point and Martin Head (see Fig. 18 for locations) supposedly predating 13.4 ka, must also be older. Sand underlying till southwest of Rockport at the head of the Bay of Fundy (Rampton and Paradis, 1981c) and marine sediments underlying and grading to glaciofluvial deltas along Bay of Fundy east of Red Head probably predate 13.4 ka. The sediments near Rockport lie on a marine bench, which has been assigned a Sangamonian age (Grant, 1980a, b; Rampton and Paradis, 1981c).

Lacustrine Deposits (Lb, Lv)

Late Wisconsinan and early Holocene glaciolacustrine and lacustrine deposits are common throughout the province. Commonly they cover small areas, are too thin to recognize, or are only present in the subsurface. Other than the sediments deposited in the lake that occupied middle and lower Saint John River valley (discussed in the section on *Undifferentiated Marine and Lacustrine Deposits*), only the lacustrine sediments deposited in Lake Madawaska cover a significant area. Lake Madawaska is the water body that occupied Madawaska and Saint John river valleys between Grand Falls and Edmundston during the Late Wisconsinan and early Holocene.

Materials and Distribution

Thin veneers of sand cover much of the terrain in most valleys draining towards retreating glacier margins. Silts and clays are also present at lower elevations. The deposits occur along the northwestern edge of the Caledonian Highlands (Rampton and Paradis, 1981c), the Clarendon Hills (Thomas et al., in press), and the Pokiok Hills; western edge of the Miramichi Highlands; and eastern edge of the Edmundston Highlands.

Varved clays underlie alluvium near the mouth of Sevogle River (see *Undifferentiated Marine and Lacustrine Deposits*; Fig. 21). Gauthier (1978a) reported at least 370 varves in a 4.3 m section in this area. Some of the surface sand in this area may also relate to glacial Lake

Figure 21. *Exposure about 4 km south of Curventon showing alluvium overlying thinly bedded brown rhythmites (6 cm thick) and varved clays.*

Sevogle (Brinsmead, 1978). Varved clays have also been noted under alluvium along Upsalquitch River (Gauthier, 1979), as many as 500 varves being noted in one section.

Horizontally interbedded sands and silty sands of probable lacustrine origin are present near Napadogan and Juniper (Fig. 22). These deposits take the form of lacustrine plains and blankets (Rampton and Paradis, 1981a). Similarly, near Sisson Branch, 0.5 to 1 m of silt and sand blanket the terrain at low elevations and a veneer of silty sand is present at higher elevations. In upper Nepisiguit River valley a thin blanket of sand (0.3 to 0.8 m thick) covers glaciofluvial deposits (glacial Lake Nepisiguit, Fig. 22). Gauthier (1982) has also mapped an area of lacustrine or low-energy sediments near the head of Bartibog River.

Glaciolacustrine deposits in valleys within the Edmundston Highlands are partially or completely covered by alluvium and colluvium. Clays occur in Three Mile Brook valley (glacial Lake Ennemond; Kiewiet de Jonge, 1951) and are also present in Ruisseau à Félix-Martin, a small valley a few kilometres to the southwest. Thick clays recorded in water well logs in Iroquois valley north of Edmundston may also be of glaciolacustrine origin.

Lake Madawaska

Beach gravels and subsurface clays abound along Saint John River valley between Grand Falls and Edmundston in the area covered by Lake Madawaska. Chalmers (1885) reported clays northwest of Grand Falls, and Kiewiet de Jonge (1951) noted varved clays just north of Grand Falls and at Van Buren, Maine. More massive clays were observed by Kiewiet de Jonge (1951) at Edmundston, and by Lougee (1954) at Van Buren, Maine. Lee (1959a) reported varved clay midway between Siegas and Grand Falls and a clay bed in an exposure along lower Green River. Kite and Borns (1980) estimated the presence of 1200 rhythmites in the latter section (Fig. 23) and reported that lacustrine rhythmites are common throughout the area. Beach gravels up to 2 m thick are present between elevations of 141 and 156 m along Saint John River valley between Grand Falls and Edmundston (Kiewiet de Jonge, 1951; Kite and Borns, 1980; Lee, 1959b). Beaches of an earlier phase of Lake Madawaska may be present at higher elevations to the west (Kiewiet de Jonge, 1951).

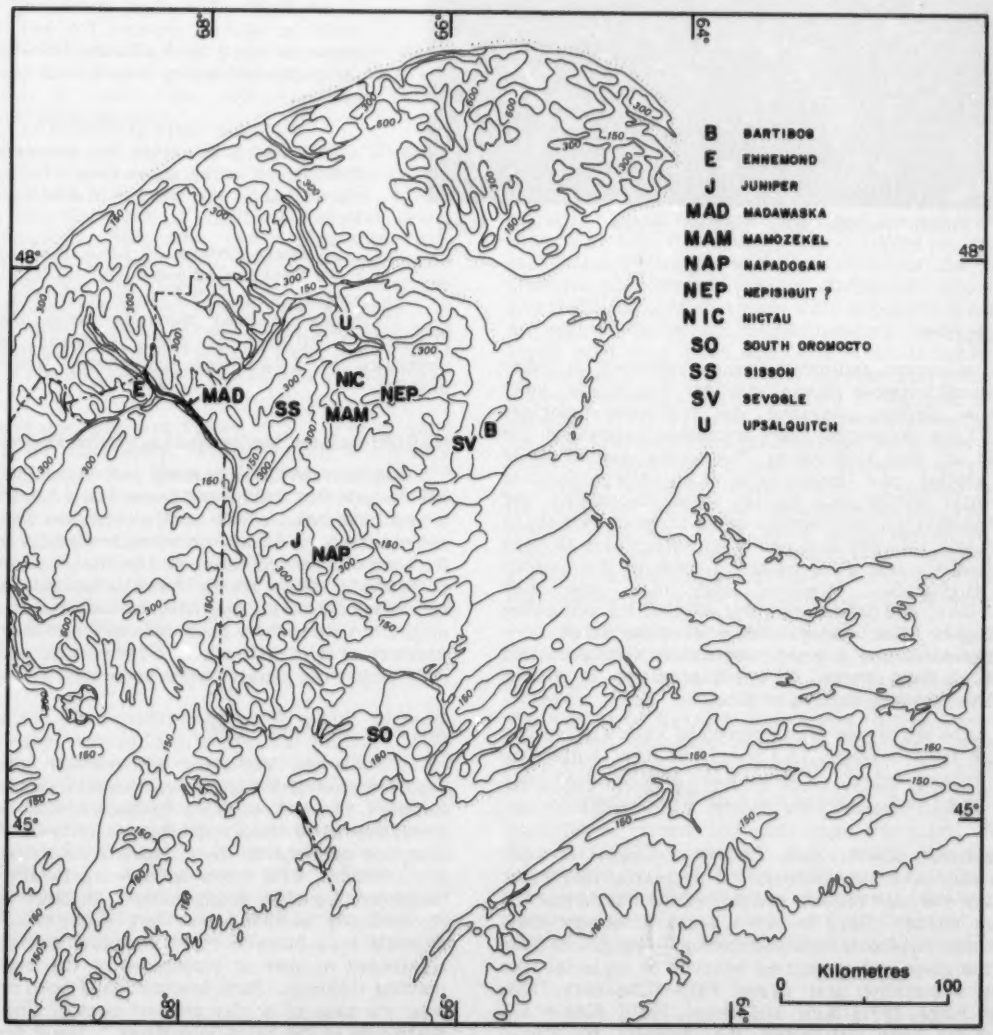


Figure 22. *Location of late Quaternary lakes in New Brunswick.*

Figure 23. *Thinly rhythmically bedded clays over till along Green River.*

damming by a remnant ice cap on the Miramichi Highlands (Gauthier, 1980a; Gauthier and Thibault, 1980). Rampton and Paradis (1981a) suggested that the sediments were deposited in fresh water at the head of a long estuary in Saint John River valley. Comparison of relative sea level curves for Saint John and for Rivière-du-Loup and Trois-Pistoles, on the Gulf of St. Lawrence 100 km north of Edmundston (Fig. 24), suggests that Lake Madawaska was the northern extension of a large, long lake (Inland Sea Acadia) occupying Saint John River valley and adjacent terrain after 12.3 ka. The curves in Figure 24 suggest that around 12 ka the Edmundston area would have been depressed about 105 m relative to sea level and that the Reversing Falls at Saint John stood about 50 m above that same level, a difference of 155 m. This difference in elevations is close to the maximum elevation of Lake Madawaska beaches.

The distribution of clay and other lacustrine sediments at Grand Falls (Kiewiet de Jonge, 1951; Lee, 1961, 1962b), including clays underlying the organics from which GSC-56 (9830 ± 160 BP, Table 3; Fig. 21) was obtained, attests to a water body with a surface elevation of 158 m covering Grand Falls. The clays underlying the latter site could be alluvial channel fill, but it hardly seems probable that a thick channel fill would develop in a short channel having drop-offs of more than 50 m at each end.

The upper clays attributed to Lake Madawaska were probably deposited in a lake restricted to Saint John River valley above Grand Falls, its level being controlled by the level of the bedrock sill at Grand Falls.

Early phases of Lake Madawaska may date from between 12.4 and 12.2 ka (cf. Fig. 20). Later stages predate 9.9 ka as indicated by radiocarbon dates on peat overlying Lake Madawaska clays (Table 3). Infilling of small remnant lakes may have continued to 8.8 ka (see *Quaternary History* for further details).

Origin and Age

Glaciolacustrine sediments were deposited in lakes (Fig. 22) formed between wasting glaciers (e.g. glacial Lake Bartibog), in valleys dammed by retreating glaciers (e.g. glacial Lake Oromocto, lakes occupying valleys on the north flank of the Caledonian Highlands, glacial Lake Ennemond, glacial Lake Upsalquitch), or in valleys dammed due to crustal tilting (e.g. glacial lakes Napadogan and Juniper). The thin glaciolacustrine sands in upper Nepisiguit River valley (glacial Lake Nepisiguit) are interpreted to have been deposited in a lake dammed by ice covering the Eastern Miramichi Highlands. However, they may have been deposited in lakes and ponds along the edge of ice occupying upper Nepisiguit River valley. Some of these lakes were probably ephemeral and drained subglacially (Rampton and Paradis, 1981c), but others, in which hundreds of varves formed, existed for long periods of time.

Most lakes within the area covered by Late Wisconsinan glaciers date from between 13.3 and 11 ka (see *Quaternary History* for further details). Glacial Lake Bartibog may have existed during an early phase of the Late Wisconsinan.

Lake Madawaska. Rhythmically bedded sediments in Lake Madawaska may have been deposited from glacial meltwater as varves, but the high organic content of the middle part of rhythmically bedded clays in some sections brings some question to this interpretation. Numerous investigators have attributed the clays and associated beaches to a glacial lake dammed by a moraine near Grand Falls (Chalmers, 1886; Kiewiet de Jonge, 1951; Kite and Borns, 1980; Kite et al., 1982). Others have attributed the deposits to glacial

Undifferentiated Marine and Lacustrine Deposits

Undifferentiated marine and lacustrine deposits have been mapped in middle and lower Saint John River valley and Grand Lake basin where both marine and lacustrine deposits are presumed to occur. Other such deposits are present along Bathurst-Curventon Valley and on the Acadian Peninsula east of Tracadie (Map 1594A). Some sandy sediments on the New Brunswick Lowlands may also be lacustrine (or estuarine) in origin, but these have been discussed in the section on marine deposits as it is difficult to differentiate possible lacustrine sediments from known marine sediments in this area.

Blankets and Plains (L.Wb)

Thick silt and clay of unknown origin have been reported to underlie areas near Fredericton and the mouth of Keswick River (Lee, 1957; Seaman, 1982b). Seaman (1982b, p. 54) described these materials as follows: "Texturally, the observed portions of these deposits consist predominantly of silt and clay, with minor sand being locally present. They range in character from massive through colour-banded to rhythmically bedded 'varve'-like deposits....At depth, these deposits may become more variable, probably containing a significant number of dropstones or till lenses deposited by melting icebergs. Such features have been recently observed near the base of a clay pit just east of Fredericton, on the north side of the Saint John River." These deposits have been

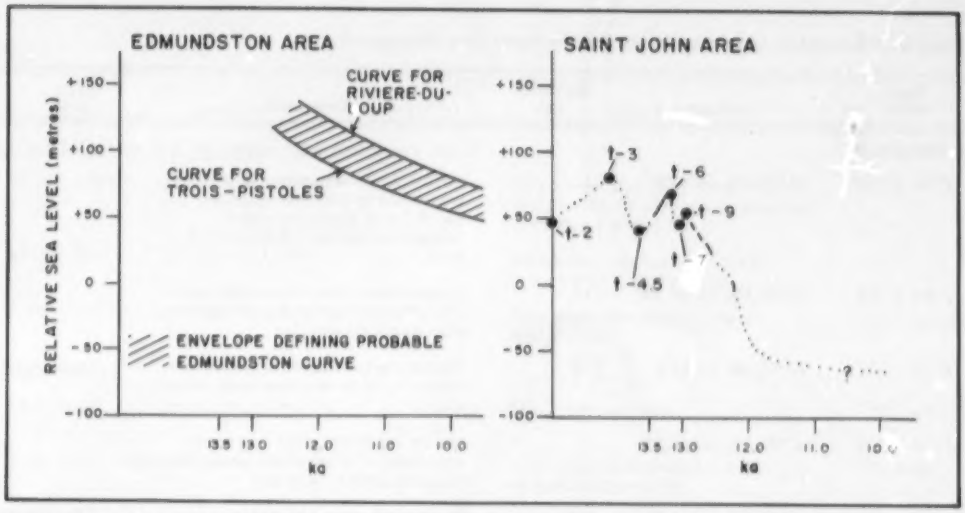


Figure 24. *Deduced relative sea level curves for the Saint John and Edmundston areas. Curves based on data in Elson (1969), Schnitker (1974), Locat (1977), Amos et al. (1980), and Figure 30.*

mapped as undifferentiated blankets and plains as have those, along lower Saint John River valley and adjacent to Grand Lake. Alluvium in Saint John valley probably covers extensive areas of these deposits.

Similar deposits of clay, silt, and sand to those described above have been reported in the subsurface from Nashwaak valley north of Fredericton (Seaman, 1981, 1982b) and along Saint John River and its tributaries between Grand Falls and Fredericton (Chalmers, 1885; Finamore, 1981; Kiewiet de Jonge, 1951; Lee, 1959b, 1962a; Rampton and Paradis, 1981a; Seaman and Thibault, 1981). Generally the clay is not obviously varved but shows sand partings. In Oromocto River valley at least 3 m of reddish clay with sand partings is present south of Blissville and more than 4 m of convoluted sandy silt and clay underlies alluvium to the north of Blissville.

Along Curventon-Bathurst Valley, 0.5 to 3 m of sand blankets much of the topography. A few thin clay or silt beds are present in the sequence. These deposits become gravelly in texture where they blanket glaciofluvial gravels. In the subsurface, varved clays (Gauthier, 1978a) are present near the mouth of Sevogle River and more massive clays near the mouth of Northwest Miramichi River valley. At one locality, 1.2 m of grey rhythmically bedded clays, which appear truly varved, is overlain by brown rhythmically bedded silts and clays with sand partings (Fig. 21).

Veneer

Thin discontinuous deposits, consisting mainly of sand, mantle much of the land surface below the limit of wave washing on higher ground where thicker undifferentiated deposits are present. Thin beds of clay or silt may occur in a few areas. Thin sand and gravel deposits noted by Rampton and Paradis (1981a) and some areas of rill wash and sheet wash noted by Lee (1962a) along Saint John valley have been included within this unit. On the steep banks of the Saint John, a loose diamicton (such as has been described by Rampton and Paradis (1981a) from near Hillman south of Woodstock) is representative of the undifferentiated deposits.

Origin

Undifferentiated marine and lacustrine deposits in Saint John River valley and adjacent areas may have been deposited in a number of environments. Early geologists (Hind, 1865; Matthew, 1879; Chalmers, 1885, 1890, 1902) considered these deposits as being of lacustrine or marine origin but could find no fossil evidence to support either origin. Kiewiet de Jonge (1951) and Lougee (1954) believed that Saint John River valley was occupied by an arm of the sea characterized by weakly saline water due to the large amounts of meltwater being discharged into the valley. Lee (1957) thought that most were deposited in a glacial lake, although he did recognize that one unit containing a marine fossil must have been deposited in an estuarine or marine environment. Finamore (1981) believed that some of the fine grained sediments were deposited in the sea; Rampton and Paradis (1981a) favoured a marine environment of deposition for fine grained sediments in middle Saint John River valley. Seaman (1982b) thought that the sediments were deposited in a complex estuarine environment. Probably both marine (or estuarine) water and lacustrine water was present in middle Saint John River valley during deposition of the fine grained sediments there. Thibault (1981) assigned most fine grained sediments in southern Oromocto River valley a glaciolacustrine origin, but some may have a marine origin.

Specifically, fine grained sediments were deposited in deep water, whereas the coarse grained sediments are the result of wave washing of higher ground (Seaman and Thibault, 1981). Sands covering fine grained sediments near the mouths of Keswick and Oromocto rivers were deposited by current or wave action as the area shoaled during Holocene isostatic crustal adjustments. Diamicton deposited on slopes along Saint John River valley probably formed as the receding water eroded steep slopes, causing the eroded material to tumble and slide down the slope. Seaman and Thibault (1981) indicated that much of the material interpreted by Lee (1957) as colluvium on slopes is material that was transported along shorelines by wave action or currents. Some gravel, sand, and silt constituting the upper part of the undifferentiated deposits is alluvium.

Table 3. Radiocarbon dates and descriptions for northern New Brunswick

Lab Dating Number	Date (BP)	Location	Elevation (m)	Comments	Reference
Late Wisconsinan Deglaciation					
GSC-3552	12 500 ± 130	47°55.5'N, 66°02'W	1	Marine shells (**Mya arenaria**?) from base of 0.4 m silty diamicton and top of 0.4 m underlying clay; sequence underlies 2 m gravel and sand	
GSC-2839	12 500 ± 360	47°57.1'N, 66°07.5'W	15	Marine shells (**Hiatella arctica**) from 3 m below surface of gravels overlying thin diamicton and clay	
GSC-1557	12 500 ± 170	47°55.1'N, 66°01'W	13	Marine shells (**Hiatella arctica**) from within 3.5 m of sand overlying silt and clay	Blake, in press
GSC-3485	12 300 ± 220	47°57.1'N, 66°07.5'W	7	Marine shells (barnacle fragments) from base of 0.2 m of diamicton overlying clay and underlying gravel	
GSC-2748	12 300 ± 210	47°49.5'N, 66°11.3'W	290	Basal gyttja from 633-638 cm below lakewater/sediment interface	R.J. Mott, pers. comm., 1983
GSC-2727	12 200 ± 230	47°59.2'N, 66°24.3'W	45	Marine shells (**Hiatella arctica**) from 2 m of gravel overlain by 0-1 m of interbedded silt and sand	
GSC-2578	12 100 ± 260	47°59.8'N, 66°39.7'W	29.5	Marine shells from 2 m of silt overlying bedrock	
GSC-2943	12 000 ± 180	47°59.4'N, 66°23.5'W	32	Marine shells (**Hiatella arctica**) from near mid-point of 16 m of gravel	
GSC-3450	11 900 ± 100	47°57.1'N, 66°07.5'W	13	Marine shells (**Hiatella arctica**) from 5 m below surface of gravels overlying thin diamicton and clay	
GSC-662	11 200 ± 200	46°54'N, 67°47'W	165	Basal gyttja from sequence of clay, gyttja, marl, and lacustrine sediments	Lowdon and Blake, 1970 Rampton and Paradis, 1981a
GSC-2804	11 100 ± 90	46°53.5'N, 67°24'W	155	Clayey gyttja from 440-446 cm below lakewater/sediment interface in Roulston Lake	Rampton and Paradis, 1981a
BGS-73	10 664 ± 77	47°27'N, 67°28'W	208	Peat from depth of 225-235 m in bog	Korpijaakko, 1976
GSC-2751	10 500 ± 150	47°37'N, 65°24'W	80	Clayey gyttja from near base of 8 m of organic lake sediments in Teagues Lake	
GSC-661	9150 ± 150	46°54'N, 67°47'W	165	Gyttja from between marl and lacustrine sediments	Lowdon and Blake, 1970 Rampton and Paradis, 1981a
Late Wisconsinan and Early Holocene Submergence and Emergence					
BGS-515	12 760 ± 170	47°40'N, 65°34'W	2	Marine shells from 1 m of clay underlying 2 m of interbedded sand and silty clay	Finamore, 1978
GSC-1383	12 600 ± 400	47°44'N, 64°56.9'W	0	Marine shells (**Mytilus edulis**) from 1 m of gravelly sand over clay	Thomas et al., 1973 Blake, in press
GSC-3284	12 600 ± 140	46°43.5'N, 64°59.6'W	8	Vertebra of balenopterid cf. **Balaenoptera acutorostrata** from 2-4 m of clay under sand veneer	
BGS-514	12 110 ± 200	47°40'N, 65°34'W	2	Marine shells from 1 m of clay underlying 0.3 m of oxidized clay and 1 m of sand	Finamore, 1978
S-1969(B)	11 830 ± 950	46°43.4'N, 64°59.6'W	8	Vertebra of balenopterid cf. **Balaenoptera acutorostrata** from 2-4 m of clay under sand veneer; original preparation (S-1969(A)) gave erroneous date	

Table 3 (cont.)

Lab Dating Number	Date (BP)	Location	Elevation (m)	Comments	Reference
Late Wisconsinan and Early Holocene Submergence and Emergence (cont.)					
GSC-2767	10 100 ± 170	47°05'N, 65°51.5'W	38	Gyttja from 253-257 cm below lake-water/sediment interface in Shaddick Lake; water level at 45 m	
GSC-2562	8710 ± 80	47°48'N, 64°33'W	5	Peat layer from base of 1 m of peat overlying sand and till	
GX-5023	8155 ± 265	47°25.5'N, 64°54.9'W	15	Peat layer from base of 2 m of peat in bog	D. Keys, pers. comm., 1983
Late Holocene Submergence					
GX-5024	4815 ± 160	49°42.1'N, 64°45.5'W	2.5 m below high tide	Basal peat over sand	D. Keys, pers. comm., 1983
GX-5022	2530 ± 215	47°43.9'N, 64°44.9'W	1 m below high tide	Wood fragments about 1 m below surface of salt marsh	D. Keys, pers. comm., 1983
Submergence of Saint John River					
I(GSC)-2	10 220 ± 350	47°19'N, 68°07.5'W	144.5	Peat from basal 1.3 cm of 0.5 m-thick peat layer overlying rhythmically bedded clay and silt and underlying sand and gravel	Lee, 1959a Kite, 1979
GSC-5	10 140 ± 150	47°19'N, 68°07.5'W	144.5	Wood? (or peat) from basal 1.3 cm of 0.5 m-thick peat layer overlying rhythmically bedded clay and silt and underlying sand and gravel	Kite, 1979
SI-3701	10 070 ± 75	47°14'N, 68°00.7'W	136	Organic silts from top of rhythmically bedded clay and silt, which are overlain by peat bed	Kite, 1979
GSC-56	9830 ± 160	47°02.4'N, 67°44.5'W	142	Gyttja about 5 m below surface and underlain by clay, till, and gravel	Dyck and Fyles, 1963 Lee, 1961
SI-3705	9720 ± 70	47°19'N, 68°07.5'W	144.5	Basal 2 cm of 0.5 m-thick peat layer overlying rhythmically bedded clay and silt and underlying sand and gravel	Kite, 1979
SI-3702	9720 ± 70	47°14'N, 68°00.7'W	136	Basal 1 cm of peat from 0.35 m-thick peat bed overlying rhythmically bedded clay and silt and underlying sand, silt, and clay	Kite, 1979
W-2927	9490 ± 350	47°14'N, 68°00.7'W	136	From 0.35 m-thick peat unit underlying 4 m of sand, silt and clay and overlying rhythmically bedded clay and silt	Kite, 1979
SI-3899A	9665 ± 75	47°19'N, 68°07.5'W	144.5	Basal 1 cm of 0.5 m-thick peat layer overlying rhythmically bedded clay and silt and underlying sand and gravel	Kite, 1979
SI-3706	9285 ± 70	47°19'N, 68°07.5'W	145	Sedge peat from near top of 0.5 m-thick peat layer overlying rhythmically bedded clay and silt and underlying sand and gravel	Kite, 1979
Alluvial Events					
GSC-18	9820 ± 130	47°13'N, 67°58'W	132	Wood within gravel layer underlying 3.5 m of fine grained sediment	Dyck and Fyles, 1963 Kite, 1979
SI-3704	8850 ± 65	47°14'N, 68°00.7'W	138.5	Wood from upper part of 0.35 m-thick peat bed underlying 4 m of sand, silt; and clay and overlying rhythmically bedded clay and silt	Kite, 1979
SI-3901A	8770 ± 100	47°14'N, 68°00.7'W	138.5	Wood, 10 cm from top of 1.2 m of organic-rich, fine grained sediment underlying 2 m of sand and silt	Kite, 1979
W-353	8250 ± 200	47°15.5'N, 68°00.7'W	135	Wood from base of 3 m of alluvium	Lee, 1959a Kite, 1979

Table 3 (cont.)

Lab Dating Number	Date (BP)	Location	Elevation (m)	Comments	Reference
Alluvial Events (cont.)					
L-190B	8200 ± 300	49°13.8'N, 67°55.7'W	143	Wood in upper part of 2 m of clay overlain by 0.6 m of silt and gravel and underlain by gravel	Lee, 1959a Kite, 1979
SI-3900	6160 ± 85	47°14'N, 68°00.7'W	138.5	Plant fragments from organic lens near top of 1.2 m of organic-rich, fine grained sediment underlying 2 m of sand and silt; some contamination	Kite, 1979
SI-3901	5795 ± 70	47°14'N, 68°00.7'W	138.5	Wood, 10 cm from top of 1.2 m of organic-rich, fine grained sediment underlying 2 m of sand and silt	Kite, 1979
SI-3703	4250 ± 50	47°14'N, 68°00.7'W	138.5	Plant fragments from organic lens near top of 1.2 m of organic-rich, fine grained sediment underlying 2 m of sand and silt; some contamination	Kite, 1979
Peat Accumulation					
BGS-521	7230 ± 110	47°37.7'N, 64°48.6'W	-2	Peat from depth of 530 cm in bog	D. Keys, pers. comm., 1983
BGS-520	3710 ± 90	47°37.7'N, 64°48.6'W	0	Peat from depth of 340 cm in bog	D. Keys, pers. comm., 1983
BGS-519	2250 ± 80	47°40.5'N, 64°45.6'W	-3.5	**Sphagnum** peat from depth of 400 cm in bog	D. Keys, pers. comm., 1983
BGS-518	250 ± 80	47°40.5'N, 64°45.6'W	0	**Sphagnum** peat from depth of 50 cm in bog	D. Keys, pers. comm., 1983
Pollen Profiles					
GSC-2872	9930 ± 160	46°53.5'N, 67°24'W	155	Silty gyttja from 406-409 cm below lakewater/sediment interface in Roulston Lake	Rampton and Paradis, 1981a
GSC-3492	8690 ± 100	47°49.5'N, 66°11.3'W	290	Gyttja from 495-500 cm below lakewater/sediment interface	R.J. Mott, pers. comm., 1983
GSC-3466	8410 ± 260	47°37'N, 65°24'W	90	Clayey gyttja from core in Teagues Lake	R.J. Mott, pers. comm., 1983
GSC-3455	8280 ± 130	46°53.5'N, 67°24'W	155	Gyttja from 350-355 cm below lakewater/sediment interface in Roulston Lake	R.J. Mott, pers. comm., 1983
BGS-76	8044 ± 71	47°26.5'N, 64°54'W	12	Peat from depth of 610-620 cm in bog	Korpijaakko, 1976
BGS-155	6630 ± 155	46°52'N?, 66°37'W	390	Peat from 2.5-2.6 m below surface of bog	Korpijaakko, 1976 Rampton and Paradis, 1981a
BGS-189B	4300 ± 130	47°04'N, 64°51'W	7	Wood from depth of 525 cm in bog	Korpijaakko, 1966
BGS-189A	3870 ± 100	47°04'N, 64°51'W	7	**Sphagnum** peat from depth of 512-525 cm in bog	Korpijaakko, 1976
Apparent Invalid Dates					
S-1969(A)	18 370 ± 500	46°43.4'N, 64°59.6'W	8	Vertebra of balenopterid cf. **Balaenoptera acutorostrata** from 2-4 m of clay under sand veneer	Rampton and Paradis, 1981b
GSC-657	13 800 ± 170	46°54'N, 67°47'W	165	Marl from near base of sequence of gyttja, marl, and lacustrine sediments	Lowdon and Blake, 1970 Rampton and Paradis, 1981b
GSC-2557	12 400 ± 110	47°20'N, 68°44.5'W	268	Basal gyttja from 835-845 cm below lakewater/sediment interface	R.J. Mott, pers. comm., 1983
SI-3899	12 160 ± 150	47°19'N, 68°07.5'W	144.5	Basal 1 cm of 0.5 m-thick peat layer overlying rhythmically bedded clay and silt and underlying sand and gravel	Kite, 1979

Undifferentiated sandy and gravelly deposits south of Bathurst were probably formed by wave washing of glaciofluvial deposits in the area. This entire area was probably covered by either a large marine or a freshwater body. As explained in the description of the origin of supposed marine sediments in this region, both possibilities exist. Varved sediments present near the mouth of Sevogle River (Brinsmead, 1978; Gauthier, 1978a) probably were deposited in a lake that was dammed by a glacier to the south. Chalmers (1888) interpreted fine grained sediments in lower Northwest Miramichi River valley as being marine, but his reasons for this interpretation are not clear.

Undifferentiated marine and lacustrine sediments on the Acadian Peninsula may have been deposited in an ice-dammed lake as glaciers retreated to the south or could be the result of wave washing by a marine or freshwater body covering much of the New Brunswick Lowlands. This part of the peninsula lies well above the maximum limit of submergence as envisaged by Gauthier (1978a), but within the extension of limits of submergence to the south proposed by Rampton and Paradis (1981b).

Age

The oldest undifferentiated deposits in Saint John River valley date from after 13 ka as the area was not deglaciated until after this time (Fig. 20). The marine (or estuarine) facies would have been deposited prior to about 12.3 ka (see *Quaternary History*), whereas the lacustrine facies was deposited subsequently, at least until 10 ka (BGS-663, -665, Table 2). Lacustrine deposition continued until uplift allowed establishment of Saint John River.

Undifferentiated marine and lacustrine deposits on the northern New Brunswick Lowlands south of Bathurst have similar ages to marine deposits mapped in that area. The older sediments probably date between 13.4 and 13 ka as deduced from Figure 20. Undifferentiated deposits on the Acadian Peninsula may be similar in age to those south of Bathurst or may be somewhat older if they relate to deglaciation of the Acadian Peninsula.

Holocene

Organic Deposits (Op)

Materials and Distribution

Organic deposits consist primarily of peat, in places muck and gyttja, and rarely marl. Silt and fine sand are interbedded in the lower part of some organic sequences. Sphagnum peat constitutes the largest part of most organic deposits in eastern New Brunswick but decreases in abundance to the west. Sedge peats of varying thickness normally underlie the sphagnum peat. Peat thicknesses are commonly more than 2.5 m and exceed 8 m in some localities.

Organic deposits underlie various peatland types including peat bogs, fens, and swamps. Domed bogs constitute the largest area of mapped organic deposits (D. Keys, personal communication, 1983). These bogs occur primarily in a broad belt extending from the southwest to the northeast of the province and includes most of the New Brunswick Lowlands. They constitute a large part of the coastal regions of the northeastern Central Lowland, Acadian Peninsula, and Acadian Islands.

Origin and Age

Most large peat bogs in New Brunswick have developed by the accumulation of organic materials on poorly drained terrain in a maritime climate. This type of bog commonly has a raised surface typical of ombrogenous peats. Other organic deposits may represent the accumulation of gyttja, muck, and woody and sedge peats in poorly drained basins and depressions.

Dated basal organic deposits throughout the province (Tables 2, 3) indicate that the accumulation of these deposits began after 10.5 ka, the accumulation being limited by deglaciation and emergence of the terrain. Radiocarbon-dated peat sequences in the northern New Brunswick Lowlands (Table 3) indicate that the upper 4 to 5 m of peat has accumulated during the last 5 ka.

Alluvial Deposits (Ap)

Materials and Distribution

Gravels, cobbly and bouldery in many places, are the main component of alluvium along the upper courses of streams. Brinsmead (1977) noted in southern New Brunswick that terrace gravels are commonly pebbly to cobbly and cleaner than glaciofluvial deposits from which they are reworked. "Downstream the alluvium generally becomes sandier and siltier, a function of finer-grained source material and lower stream gradients. Most alluvial gravel and sand has a thin cover of silt or silty sand." (Rampton and Paradis, 1981c, p. 12). Major rivers bordered by coarse alluvial terraces and floodplains include the Magaguadavic, upper reaches of streams draining the Caledonian Highlands, Little Southwest Miramichi, Renous, Dungarvon, Bartholomew, and Nashwaak (Rampton and Paradis, 1981a, b), and the Nepisiguit, Northwest Miramichi, Upsalquitch, Kedgwick, Restigouche, Tobique, and Mamozekel (Gauthier, 1982, 1983). Major rivers crossing the New Brunswick Lowlands are bordered by sandy alluvium with local gravel beds (Rampton and Paradis, 1981b). Madawaska River valley is filled with alluvium consisting mainly of pebbly sand.

Alluvial deposits are commonly underlain by glaciofluvial, lacustrine, or marine deposits. Lacustrine and marine clays are common under alluvium below limits of marine or lacustrine submergence along lower Keswick and Nashwaak (Rampton and Paradis, 1981a; Seaman, 1981, 1982b), Northwest Miramichi and Sevogle (Brinsmead, 1978), and Upsalquitch valleys (Gauthier, 1979). Alluvium is also commonly interbedded with lacustrine deposits along the lower course of tributaries to Saint John River between Grand Falls and Edmundston (Gauthier, 1983).

Fine grained alluvium (fine sand, silt, and clay with variable organic contents) is dominant along the lower courses of stream valleys where they approach sea level, particularly Kennebecasis River and Belleisle Creek northeast of Saint John (Rampton and Paradis, 1981c). Marine clay and silt is commonly interbedded with alluvium where streams debouch into the ocean.

Alluvial deposits form terraces and floodplains along streams; few streams are not bordered by narrow terraces or floodplains. Stream-cut scarps, channel traces, oxbows, levees, and point bars characterize the surface of terraces and floodplains.

Saint John River Valley

Saint John River valley above Edmundston is characterized by coarse grained alluvium, mainly sandy and gravelly, but bouldery in a few locations. Gravel and sand are interbedded with clay and silt between Edmundston and Grand Falls. Gravels and sands, 3 to 4 m thick, constitute channel sediments; clays and silts, 1 to 5 m thick, constitute channel fills; and sand and silty sand, up to 7 m thick, constitute the overbank alluvium along this segment of the river (Kite and Borns, 1980).

Below Grand Falls, extensive narrow terraces are present to near the mouth of Keswick River. Lee (1962a, p. 5) described the terrace deposits as follows: "The deposits have a vertical change in texture from clean, well-sorted, gravels and sands of the terrace deposits near the surface, to underlying, moderately sorted and more crudely bedded, valley-train gravels and sands. Beds of the terrace deposits generally are nearly horizontal, but dip slightly downstream." The alluvium is generally gravel as described by Lee, but low terraces, especially south of Woodstock, are composed largely of sand and silty sand (Chalmers, 1885). The bedding may be locally steep in the form of foreset beds, and the alluvium commonly overlies lacustrine and (or) marine clay and silt (Rampton and Paradis, 1981a).

Below the mouth of Keswick River the alluvium consists of sand and silt with a few gravel and clay beds. It is underlain by lacustrine and (or) marine clays. Commonly, the alluvium may be highly organic.

Origin and Age

Alluvium along most streams is the result of erosion of the surrounding areas by streams and redeposition of the eroded detritus along the stream course. Much alluvium consists of reworked colluvial, glaciofluvial, fluvial, lacustrine, and marine sediments as unconsolidated materials are most easily eroded. Most coarse textured alluvium is probably the result of preservation of channel bars, point bars, and channel-bottom sediments. Sandy and silty alluvium is generally deposited as levee or overbank floodplain deposits (Kite et al., 1982).

Few radiocarbon dates are available on terrace deposits other than those adjacent to Saint John River. Generally these postdate Late Wisconsinan deglaciation, and along their lower courses postdate Late Wisconsinan or Holocene emergence. Floodplain deposits are being actively formed today.

Saint John and Madawaska Rivers. Alluvium along Madawaska and Saint John rivers between Edmundston and Grand Falls was deposited as an infilling of Lake Madawaska. A complex of coarse and fine grained sediments with thick channel fills was deposited in this low-energy, slack water environment. The sandy alluvium filling Madawaska valley north of Edmundston was apparently deposited during spring flooding by the upper Saint John as it graded to an intermediate level of Lake Madawaska (Kiewiet de Jonge, 1951). Northward-dipping crossbedding has been noted by Kiewiet de Jonge (1951) and Lee (1955) in the upper part of the Madawaska alluvial sand plain.

Alluvium below Grand Falls was deposited in the form of a long time-transgressive delta by a river grading to a regressing water body (Rampton and Paradis, 1981a); foresets in the alluvial sequence indicate infilling of depressions as the delta front progressed downstream. Kiewiet de Jonge (1951) recognized that these deltaic deposits might be difficult to differentiate from glaciomarine deltas and kame deltas deposited earlier along Saint John River valley.

Alluvial deposition appears to have begun around 9 ka between Edmundston and Grand Falls (Kite et al., 1982) and continued until at least 4250 ± 50 BP (SI-3703, Table 3), forming a continuous high terrace, the Seigas Terrace (Kite and Borns, 1980), along this part of the Saint John. Lower terraces were then formed as the river incised to its present level.

Alluvium in the high terraces downstream from Grand Falls postdates final Holocene emergence of this area. With the exception of a date on floodplain alluvium of 360 ± 80 BP (BGS-664, Table 2), no dates are available on alluvium below Grand Falls.

Marine Deposits

Intertidal Plains and Salt Marshes (Wp)

Interbedded peat, muck, silt, and sand constitute the deposits under salt marshes and intertidal flats. These deposits are most common at the head of Chignecto Bay, although small salt marshes and intertidal flats can be found along most of the coastline in areas protected from wave action by baymouth bars, beaches, and spits. The thickest intertidal deposits can be found in the Tantramar, Memramcook, Petitcodiac, Hopewell, and Shepody marshes. Rampton and Paradis (1981b, p. 14) have described them as follows: "Their poorly drained surfaces are relatively flat, and only interrupted by lakes and tidal channels. Presently, most salt marshes are protected from flooding by dykes and dams. The intertidal deposits are mainly silty clays, although thin peaty beds can be present throughout their total thickness. Those along the Petitcodiac River have been described as having a texture that 'varies from clayey silt to sandy silt, these textural variations often being found rhythmically interstratified.' (Lohse, 1974c). The organic content of these sediments is variable, probably a function of the rate of accumulation of the intertidal deposits and rate of coastal submergence. The intertidal sediments are frequently more than 10 m thick along those estuaries draining to Chignecto Bay."

Marine Beaches, Bars, and Spits (Wr)

Most bars, beaches, and spits consist of sand and pebbly sand along the eastern coast (Rampton and Paradis, 1981b) and pebbly or sandy gravel along the Fundy coast (Rampton and Paradis, 1981c). Driftwood commonly litters these features. Near Shediac the marine sediments are about 1 m in average thickness (Lohse, 1974b) but are thicker where sand dunes have developed. In some localities, intertidal sediments underlie the beach sands.

Small storm ridges and windblown sand dunes are associated with bars, beaches, and spits (cf. Airphoto Analysis Associates Consultants Limited, 1975a; Hunter and Associates, 1982). Spectacular sand dunes are associated with beaches on the northwestern edge of Miscou Island.

Large baymouth bars and spits characterize the eastern coastline of the New Brunswick Lowlands between Cape Tormentine and Caraquet; major spits are also present at Bathurst. In other areas the beaches are narrow and baymouth bars and spits are present only adjacent to mouths of streams because of steep rocky shorelines, high tidal ranges, or limited fetch.

Origin and Age

Except where protected by dykes, most intertidal plains and salt marshes are actively forming at present. Some of these deposits may be as old as 4 ka as intertidal muds cover tree stumps that date from this interval in Tantramar Marshes (Fig. 19, Table 2).

Beaches, bars, and spits are essentially modern features that continue to be reworked as coastal recession continues. Some stabilized sand dunes associated with the beaches at the north end of Miscou Island may date from the 16th Century, based on the position of whale bones killed by early explorers (Goldthwait, 1912, 1914).

Eolian Deposits

Apart from thin, unmapped silt deposits of a possible loessial origin, the only eolian deposits in the province are a few sand dunes near the mouth of Oromocto River, a few sandy ridges on the undifferentiated marine and lacustrine deposits at the mouth of Keswick River, sand dunes

associated with beaches and spits, and a few clifftop dunes. The sand dunes at the mouth of the Oromocto are up to 20 m high (Seaman and Thibault, 1981); the supposed dunes at the mouth of the Keswick are less than 2 m high; and those along the coastline north of the Miramichi estuary are 3 to 4 m high. Eolian deposits are not shown as a unit on Map 1594A, but the main areas of sand dunes are indicated by symbols.

The Oromocto and Keswick dunes are stabilized except for rare blowouts (Seaman and Thibault, 1981) and probably were formed immediately following recession of the water body occupying this part of Saint John River valley. Sand dunes associated with sea cliffs, beaches, bars, and spits are continuously being formed as these features follow the receding coast line and as such are relatively youthful. Some of the sand dunes on northern Miscou Island may relate to Holocene emergence of this area and thus may predate 10 ka.

Quaternary

Colluvium (Cb)

Materials and Distribution

Colluvium consists primarily of weathered and disintegrated local bedrock moving downslope through mass wastage. Generally, it thickens and interfingers with alluvium at the base of slopes. In many places, colluvium is composed in part of reworked till moving downslope through creep. Locally, it may consist of talus from rock falls off steep bedrock exposures (Gauthier, 1982, 1983).

Colluvium has only been mapped along the northern edge of the Central Plateau of the Caledonian Highlands, although it is present in small amounts throughout the province. Colluvium covers steep and moderately steep slopes throughout the province, but the small areal extent of such deposits prevents delineation on Map 1594A. Most major rivers such as the Saint John, Patapedia, Restigouche, Upsalquitch, Nepisiguit, Miramichi, and Tobique have colluvial blankets on slopes forming their valley walls where the streams are incised (cf. Gauthier, 1982) and where stream erosion is not continuously exposing bedrock.

Origin and Age

Colluvium is produced by the downslope movement of material by creep and slope wash. Solifluction also may have been an active process in high areas not covered by Late Wisconsinan glaciers.

Colluvium postdates the glaciation of any landscape on which it is present. Thus, most colluvium is post-Late Wisconsinan in New Brunswick. Older colluvium is possibly present on the Caledonian Highlands and Northern Miramichi Highlands, which may only have been glaciated prior to the Late Wisconsinan. Colluvium along actively incising and eroding rivers and coastlines dates from the late Holocene.

QUATERNARY HISTORY

The Quaternary history of New Brunswick can be separated into three intervals: (1) the period predating the Late Wisconsinan (prior to about 20 ka) during which only fragmentary evidence is available for documentation and dating of events, (2) the Late Wisconsinan (approximately 20-10 ka ago) during which major fluctuations in glaciers, and water and land levels are discernable, and (3) the Holocene (10 ka until present), devoid of glacial events. Certain events, which occurred over a period of time, transcend these rather arbitrary time intervals and will be discussed in view of this. The Late Wisconsinan has been separated into phases based on the relative strength of ice centres and on the ice flow patterns that they induced. These are defined in detail in the section discussing the *Late Wisconsinan*.

Pre-Late Wisconsinan

Landscape Development

As discussed in the section dealing with physiography, most major valleys and landforms were formed prior to the Quaternary. Possibly, however, continued uplift of the landscape has resulted in the eroding and deepening of valleys during the Quaternary. Some drainage diversions have resulted due to the repeated glaciation and submergence of the landscape during the Quaternary. Notable are the wind gaps present in the Northern Miramichi Highlands through which large streams must have previously flowed; and the southern course of Saint John River downstream of Grand Lake and the southeastward course of Petitcodiac River downstream of Moncton, both of which appear to have been diverted from earlier northeastwardly courses to Northumberland Strait. These courses all were probably developed during glaciations predating the Wisconsinan.

Nonglacial Intervals

Weathered bedrock and regolith, in both the surface and the subsurface on the Northern Miramichi Highlands, have been partly attributed to pre-Wisconsinan weathering (Chalmers, 1887a, 1888; Lee, 1959a, 1962a; Gauthier, 1980b, 1983; Wang et al., 1981) and were probably developed during a nonglacial interval. Parts of the Northern Miramichi Highlands have a cover of weathered granite and regolith containing clay minerals that probably formed in a climate similar to today's over a long period of time (Wang et al., 1981). This suggests that such surfaces have been subjected to long periods of weathering in a nonglacial climate.

Glacial Intervals

Parts of the Northern Miramichi Highlands and Caledonian Highlands were probably glaciated only prior to the Late Wisconsinan, each highland possibly having been glaciated during different time intervals. Striae, boulder trains, and stratigraphic units that were probably formed prior to the Late Wisconsinan have also been identified in other areas of New Brunswick.

Northern Miramichi Highlands

The last glaciation of the high parts of the Northern Miramichi Highlands probably predates the Sangamonian. The occurrence of till on Big Bald Mountain (Gauthier, 1980b) and of granitic erratics, presumably from sources to the west, on mounts Denys and Fronsac (Gauthier, 1979) suggests that the area has been glaciated. The subsequent development of the deeply weathered granite, tors, nivation terraces, sorted polygons, and thick colluvium suggests that this glaciation was of some antiquity, likely predating the Sangamonian.

The tors, nivation terraces, sorted polygons, and thick colluvium present on the Northern Miramichi Highlands suggest a long interval of exposure to mean annual temperatures well below freezing. Tor development at northern latitudes is generally attributed to the parallel retreat of scarps near hill crests over a long period of time in a periglacial environment (Demek, 1969; Hughes et al., 1972). Thus the tors on the Northern Miramichi Highlands suggest that the area may have been free of ice for at least one major glaciation. The cirque-like features present in the area (Gauthier, 1979b) also suggest some antiquity of the landscape; modification of cirques to their present subdued cirque-like forms would require a long period of time. Gauthier (1979, 1983) has suggested that the Northern Miramichi Highlands either may have been covered by cold-based, nonerosive glacial ice during the Late Wisconsinan or may only have been covered by earlier glaciers

(Gauthier, 1980b). The presence of periglacial landforms makes the latter hypothesis more plausible.

Striae on intensely polished rocks along the eastern edge of the Northern Miramichi Highlands likely were formed by ice flowing from the west. It seems doubtful that the surface of the highlands would have been unaffected by glacial action (whether it be by cold- or warm-based glaciers) if the highlands were overtopped by ice flowing from farther to the west. Ice passing from west to east over the highlands would have had a higher velocity over the crest of the highlands than on their eastern flank, given a constant volume of ice moving across each area (Fig. 25), and should have resulted in greater erosion over the crest of the highlands than on their flanks. This would have removed the tors, nivation terraces, and other periglacial features from the highlands. More probably the ice that inscribed the striae on the eastern edge of the Northern Miramichi Highlands originated in small ice fields along the eastern edge of the highlands, and the crest of the highlands was unaffected.

Surrounding those parts of the Northern Miramichi Highlands that were likely glaciated during the pre-Wisconsinan is a belt of terrain that lacks tors and other obvious periglacial features but is characterized by highly oxidized soils and ablation moraines with muted topography, the largest moraine being at Nepisiguit Lakes (Map 1594A). This topography is not typical of that resulting from Late Wisconsinan glaciation. The topographic modification and weathering suggests an Early or Middle Wisconsinan age for this terrain, although an even earlier glaciation cannot be dismissed.

Caledonian Highlands

The Central Plateau of the Caledonian Highlands is drift mantled and exhibits a muted glacial topography. However, terrain surrounding the Central Plateau, especially around its northern periphery, is deeply dissected by intricate dendritic stream patterns (Fig. 26; see also Rampton and Paradis, 1981c). Deeply weathered bedrock, intensely oxidized soils, and sorted polygons in some glaciofluvial deposits are also present at some localities. These phenomena suggest that some time has passed since the Caledonian Highlands were glaciated. The Caledonia glacial phase, responsible for the last glaciation of the Central Plateau (Rampton and Paradis, 1981c), has been assigned an Early or Middle Wisconsinan age because striae trending in directions similar to those of the Caledonia flow pattern (Map 1594A, Fig. 27) are present on a marine rock bench (Fig. 28), which is believed to be Sangamonian in age (Grant, 1980a,b). The erosion and drainage pattern along the flanks of the Central Plateau is deemed to have developed during Middle to Late Wisconsinan and Holocene time.

During the Caledonia phase, southeastwardly flowing glaciers appear to have dominated. Rampton and Paradis (1981c, p. 18) stated: "During most of the Caledonia phase, it appears that ice was flowing from a source area to the northwest across the Caledonia Highlands as attested to by striae generally showing a southeast movement of ice. Fragments of Mississippian conglomerate and sandstone in the drift capping the uplands are also indicative of such glacier flow. However, some indication of northward flow is given by the presence of Precambrian metavolcanic erratics north of the northern limit of bedrock outcrops of this type .... Although, rocks from the Caledonian Highlands could have been transported northward by streams, the erratics ... were over 0.3 m diameter and found at interfluve localities. Chalmers (1890) has indicated that ice moved both northward and southward in the vicinity of Gowland Mountain. However, his direction of ice movement appears to be based on the aspect of outcrops showing glaciated and unglaciated surfaces (Chalmers, 1890, p. 45N)". The exact source of ice during the Caledonia phase is uncertain, although the flow pattern suggests a source farther to the northwest, possibly the Gaspereau Ice Centre (Fig. 28). It is unclear whether glaciers that glaciated the Central Plateau continued southward to override weathered bedrock along the north shore of the Minas Channel in Nova Scotia (Stea, 1983) or stagnated along their northern edge. A possible limit has been drawn on Figure 28.

During the Caledonia phase one major area of ribbed moraine was formed (near McFadden Lake) and eskers were formed in a few valleys (Rampton and Paradis, 1981c; Map 1594A). Short-lived lakes may have formed along the northwestern edge of the Central Plateau. Walton Lake valley south of Chalmers Settlement probably served as a spillway for water draining to the Bay of Fundy, where a delta formed at the mouth of Salmon River.

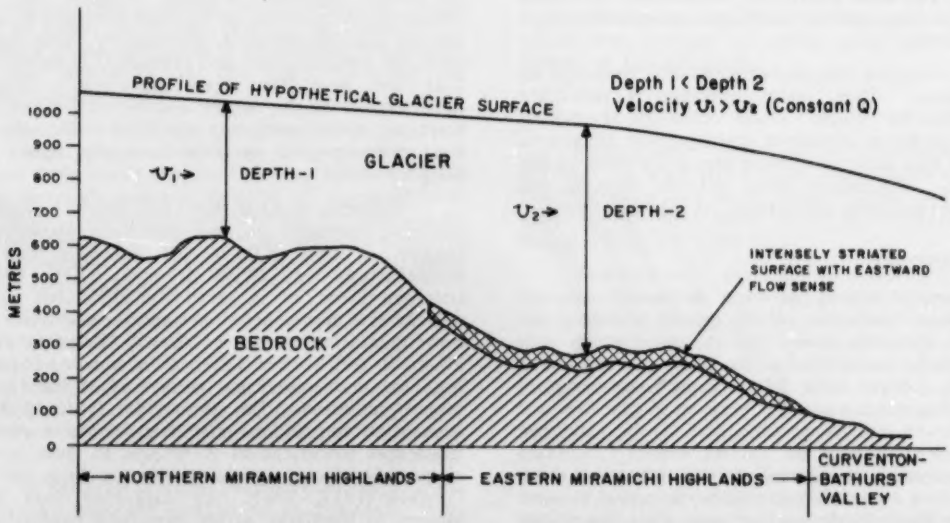


Figure 25. *Profile of Miramichi Highlands showing physics of hypothetical eastward-flowing glacier. See text for discussion.*

Figure 26. *Dissection and drainage patterns around the northwest edge of the Central Plateau of the Caledonian Highlands east of Sussex. Approximate limit of Late Wisconsinan glaciation marks the northwest edge of deep dissection and dendritic drainage (NAPL A21331-62, 63, 64).*

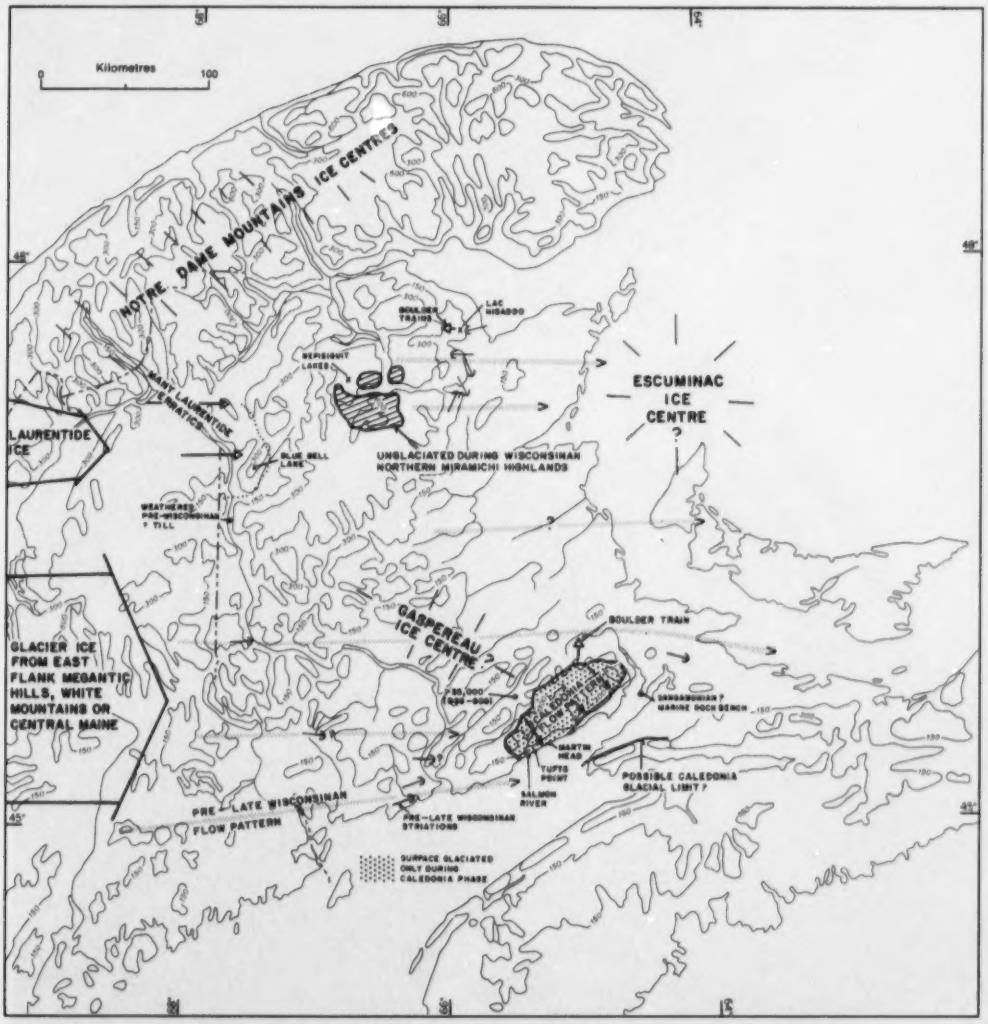
A narrow belt of terrain, along the northern flank of the area glaciated during the Caledonia phase, exists that shows few characteristics of adjacent terrain glaciated by Late Wisconsinan glaciers. This terrain was probably glaciated after the Caledonia phase but before the main Late Wisconsinan advances.

Other Glacial Indicators

Striae and glacial drift that may also record pre-Late Wisconsinan glacial events are present in localities other than the Northern Miramichi Highlands and Central Plateau.

Striae. A summary of striation patterns is shown on the back of Map 1594A. At a number of localities in the Bathurst area northeastward, south-southeastward, and westward trending striae occur on oxidized rock surfaces, where eastward trending striae are present on unoxidized surfaces (Gauthier and Cormier, 1977). This indicates that the ice movements producing the striae on the oxidized surfaces were separated from the later eastward ice flow of Late Wisconsinan age by an interval of deglaciation, rock exposure, and weathering. These older striae must relate to pre-Late Wisconsinan or possibly pre-Wisconsinan glaciations. The westward trending striae suggest movement from an area to the east where a major ice centre (Escuminac) developed during the Late Wisconsinan; the south-southeastward trends indicate a source in the Notre Dame Mountains; and the northeastward trends, a source in the Central Lowland where the Gaspereau Ice Centre developed during the Late Wisconsinan (Fig. 27).

In other parts of New Brunswick, eastward trending striae are present that predate the formation of known Late Wisconsinan striae (Striation Summary on Map 1594A; Fig. 28). At some localities the eastward trending striae appear to be present on oxidized surfaces protected from subsequent glacial polish. In addition, some of the striae assigned to the Kent flow pattern may actually be older. The distribution of the eastward trending striae suggests regional glaciation with movement to the east out of central Maine (Fig. 28). The absence of Laurentide erratics on the Miramichi Highlands (Gauthier, 1983), Caledonian Highlands, New Brunswick Lowlands (Chalmers, 1895), and over most of the St. Croix Highlands (cf. Thomas et al., in press) precludes a Laurentide source. Chalmers (1895) also favoured the east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NOTRE DAME MOUNTAINS ICE CENTRES
LAURENTIDE ICE
ESCUMINAC ICE CENTRE ?
GASPEREAU ICE CENTRE ?
UNGLACIATED DURING WISCONSINAN NORTHERN MIRAMICHI HIGHLANDS
GLACIER ICE FROM EAST FLANK MEGANTIC HILLS, WHITE MOUNTAINS OR CENTRAL MAINE
PRE-LATE WISCONSINAN FLOW PATTERN
BOULDER TRAIN
POSSIBLE CALEDONIA GLACIAL LIMIT ?
SURFACE GLACIATED ONLY DURING CALEDONIA PHASE

Figure 28. *Selected pre-Late Wisconsinan glacial phenomena.*

side of the Appalachians, especially the Notre Dame Mountains, as a source of ice for glaciation of eastern New Brunswick. The eastward trending striae date from a glaciation or glaciations predating the Late Wisconsinan. Rampton and Paradis (1981b, c) have previously assigned some of the striae to the Caledonia phase, but analysis of the probable source of the glacier ice producing the eastward trending striae and the Caledonia flow pattern places this assignment in doubt.

Boulder Trains. A number of boulder trains occur in New Brunswick suggesting ice flow directions that do not fit the flow patterns deduced for the Late Wisconsinan (Fig. 27, 28). These movements include the northward movement of Precambrian metavolcanic boulders from the Central Plateau of the Caledonian Highlands (Rampton and Paradis, 1981c) and the westward and northern movement of granitic boulders from the Nigadoo Lake intrusion on the Eastern Miramichi Highlands (Gauthier, 1978a). The former predates dissection of the Caledonia Highlands during the Wisconsinan and the latter correlates with pre-Late Wisconsinan striation patterns.

Laurentide erratics in New Brunswick are confined mainly to Saint John River valley and adjacent areas west of Plaster Rock and north of the Kintore Hills (Fig. 28). South of the Kintore Hills they are rare. The Laurentide erratics (mainly gneisses and granulites) are most abundant in the glaciofluvial deposits underlying the surface till. This stratigraphic relationship and the general distribution of Laurentide erratics suggest eastward movement of Laurentide ice around the northern end of the Mégantic Hills in eastern Quebec. This glacial movement probably occurred prior to the Late Wisconsinan or, at the latest, during its early part.

Kite et al. (1982) have assigned a Late Wisconsinan age to a dense grey till containing Laurentide erratics underlying a brown surface till in northern Maine. They believed that the two tills constitute a drift deposited during a single period of glacial cover. In northern New Brunswick, however, grey and brown tills are separated by fluvial (glaciofluvial?) sediments at some localities (Rampton and Paradis, 1981a) and indicate a possible nonglacial interval between the glaciations responsible for the two tills. Deglaciation of northern New Brunswick and Maine may have only occurred at the eastern extent of Laurentide ice in New Brunswick prior to the development of Late Wisconsinan regional ice caps over both New Brunswick and Maine. The large volume of subtill glaciofluvial deposits present along Saint John River and towards Plaster Rock (Map 1594A), especially the thick sequence of eastward-dipping glaciofluvial sediments near Blue Bell Lake (Fig. 28), suggest fluctuation of an ice margin in this area.

Stratigraphy. Identified pre-Wisconsinan glacial and nonglacial deposits underlying Late Wisconsinan deposits, exclusive of the weathered bedrock common to much of the province, are present in Saint John River valley above the mouth of Nackawic River, in the Anagance Ridges, and possibly along the Bay of Fundy.

In Saint John valley interbedded brown and grey tills, commonly compact, are present. This may reflect changes in modes of glacial flow (Lee, 1962a), changes in ice sources (Rampton and Paradis, 1981a), or the difference in the effects of oxidation on ablation and lodgment tills. The presence of glaciofluvial (or alluvial) sediments between compact (presumably lodgment) tills at some localities, however, suggests deglaciation (interstadial or interglacial) of Saint John River valley. In a section south of Grand Falls a till, containing Laurentide erratics, having an east-west till fabric and underlying a thick sequence of glaciofluvial sediments and multiple tills (Rampton and Paradis, 1981a) is oxidized to a depth of 1 m (A of Fig. 10). This suggests a period of interglacial weathering and a pre-Wisconsinan age for the till.

Interlayered till, clay, and sand containing organic layers underlies till near Sussex (Fig. 19); the organic material was dated at >35 000 BP (BGS-806, Table 2). Gravel and sand underlying the dated material is oxidized and contains granitic clasts, which are highly weathered. This evidence points to the gravel, sand, and organic materials being deposited during an interglacial interval and being overridden during the Late Wisconsinan. Gravel and sand at some localities north of Sussex to Cornhill along Smith Creek is oxidized similarly to the gravel and sand at the site near Sussex. The gravel and sand deposits along Smith Creek are usually capped by till (cf. Brinsmead, 1977).

Along the Bay of Fundy, complex sequences of marine and glacial sediments are present at Salmon River, Tufts Point, and Martin Head (Fig. 28). Angular gravel and silty sand underlying till (Caledonia phase) in a section near the mouth of Salmon River probably relate to Middle or Early Wisconsinan high sea levels as do organic-rich (?) diamictons underlying till to the east at Tufts Point. Thick, fine red sand at Tufts Point underlying coarse crossbedded sand and gravel (Late Wisconsinan age?) is also probably Middle or Early Wisconsinan in age. Silt and sand with clayey beds underlying gravel and diamicton at Martin Head may relate to the same high relative sea level.

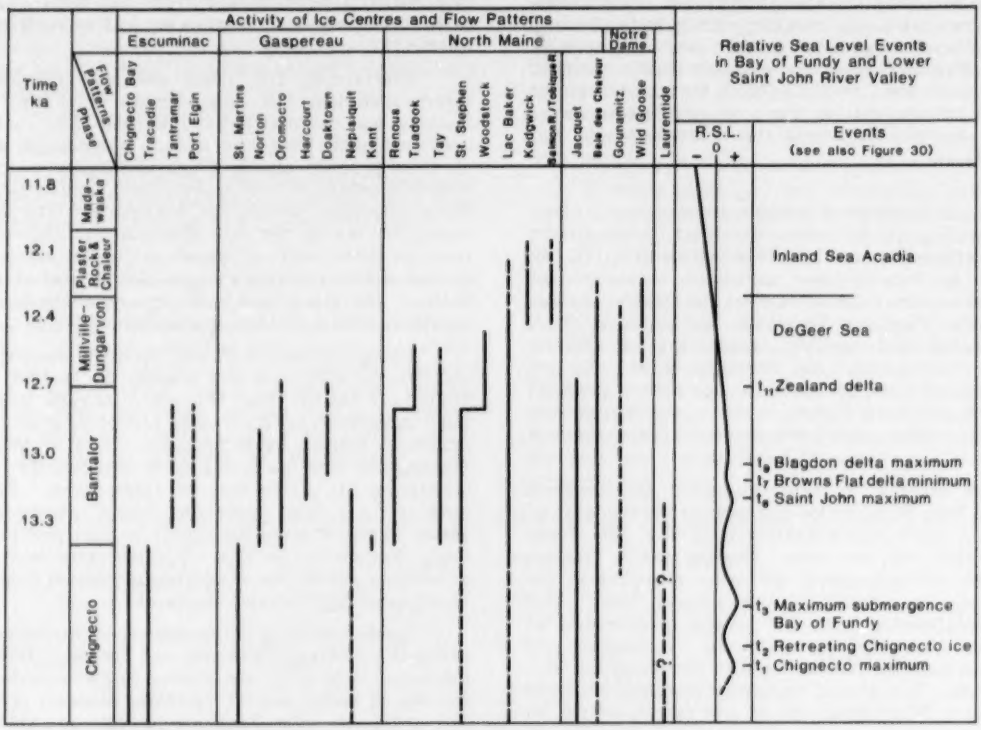
Sand underlying till southwest of Rockport at the head of the Bay of Fundy (Rampton and Paradis, 1981c) and marine sediments underlying and grading to glaciofluvial deltas along the Bay of Fundy east of Red Head probably predate 13.4 ka. The sediments near Rockport lie on a marine bench, which has been assigned a Sangamonian age (Grant, 1980a,b; Rampton and Paradis, 1981c).

Subtill gravels are present on the eastern part of the Acadian Peninsula around Tracadie (Gauthier, 1983). These gravels commonly contain a large number of clasts with lithologies common to the Miramichi Highlands and Jacquet Plateau and were either deposited by interglacial streams prior to the formation of the Curventon-Bathurst Valley or more likely were deposited by glacial meltwater from glaciers flowing from the Miramichi Highlands, possibly during the glacial interval when eastward ice flow occurred in New Brunswick. As indicated in the section on striae, this probably occurred prior to the Late Wisconsinan.

Late Wisconsinan

The Late Wisconsinan in New Brunswick can be divided into six phases (Table 4): 1) the *Chignecto phase* (Rampton and Paradis, 1981b, c) during which the Escuminac Ice Centre strongly influenced flow patterns in eastern New Brunswick; 2) the *Bantalor phase* (includes Bantalor, Kent, Sackville, and Tormentine phases of Rampton and Paradis, 1981a, b, c) during which the Gaspereau Ice Centre influenced flow patterns over most of eastern New Brunswick and the Escuminac Ice Centre only affected flow patterns in the extreme eastern part; 3) the *Millville/Dungarvon phase* (Rampton and Paradis, 1981a, b) during which ice ceased to flow effectively across the Southern Miramichi Highlands and glacial flow was confined mainly to terrain adjacent Saint John River valley in western New Brunswick; 4) the *Plaster Rock phase*; 5) the *Chaleur phase* (Gauthier, 1983; Rampton and Paradis, 1981a) marked by disintegration of the Northern Maine Ice Divide and by a glacial readvance in Baie des Chaleurs; and 6) the *Madawaska phase* during which Lake Madawaska persisted in Saint John River valley above Grand Falls.

Table 4. Summary of glacial flow patterns and relative sea level during the Late Wisconsinan.



Glacial flow patterns (Fig. 27) have been established for the first five phases from striae, flutings, and drumlins and from glacier gradients deduced from esker systems. "Generally each flow pattern has a distinctive areal and relative chronologic character that suggests one or more centre of ice accumulation and dispersion." (Rampton and Paradis, 1981a, p. 19).

Chignecto Phase

Ice Flow Patterns

During the Chignecto phase an ice cap centred near the north end of Prince Edward Island, the Escuminac Ice Centre (Fig. 29), expanded causing ice to flow across the Chignecto isthmus down Chignecto Bay past Alma (Chignecto flow pattern, 4 of Fig. 27) and northwestward across the Acadian Peninsula (Tracadie flow pattern, 5 of Fig. 27). Red fine grained sediments were incorporated into the glacier as it flowed down Chignecto Bay resulting in a red silty till being deposited over gravels rich in Caledonian clasts at Fundy National Park. On the northern edge of the Acadian Peninsula ice was moving eastward down Chaleur Bay (early Jacquet flow pattern, 2 of Fig. 27).

Ice moving southeast across the southwestern end of the Caledonian Highlands (St. Martins flow pattern, 3 of Fig. 27) has been assigned to the Chignecto phase (Rampton and Paradis, 1981c) and indicates an active Gaspereau Ice Centre during this glacial episode. Flow from the northern and eastern flanks of the Gaspereau Ice Centre was diverted down Curventon-Bathurst Valley by Escuminac ice. Supposedly the North Maine Ice Divide and ice caps in the Notre Dame Mountains were active, preventing intrusion of Laurentide ice.

Local Glacial Events

Acadian Peninsula. The complex interaction between eastward flowing ice and northwestward flowing ice on the Acadian Peninsula is reflected not only by the eastward and westward dispersion of clasts from a diabase dyke running the length of the Acadian Peninsula (Gauthier, 1978a), but by a glacial meltwater pattern along the Paquetville interlobate moraine (1 of Fig. 20; Map 1594A) during deglaciation. The westward convexity of morainic ridges at Allardville (2 of Fig. 20) suggests that the ice retreated from the western edge of the ridges prior to their deposition. At several localities along the length of the Paquetville moraine, however, drainage appears to have been blocked by ice on both the southeast side and the northwest side during its formation. Water was ponded in low areas along the axis of the peninsula and diverted northeastward along meltwater channels, which virtually parallel the modern drainage divide. A sandy delta-like feature below 15 m elevation on Pollard Brook, just southeast of Paquetville, may have been graded to sea level at this time.

During retreat of Escuminac ice an ice-thrust moraine, as evidenced by contorted glaciofluvial sediments, was constructed north of Bois-Gagnon (Map 1594A, Fig. 20). During further retreat of Escuminac ice a short-lived glacially dammed lake may have formed as the ice continued to block the streams draining southeastward on the Acadian Peninsula.

Bay of Fundy. As Escuminac ice retreated across the southwestern end of the Caledonian Highlands and up Chignecto Bay, deltas formed where glacier meltwater reached the sea. Near St. Martins deltas formed at about 48 m elevation (Fig. 18) until the ice retreated across the crest of the highlands and the sediment source was cut off.

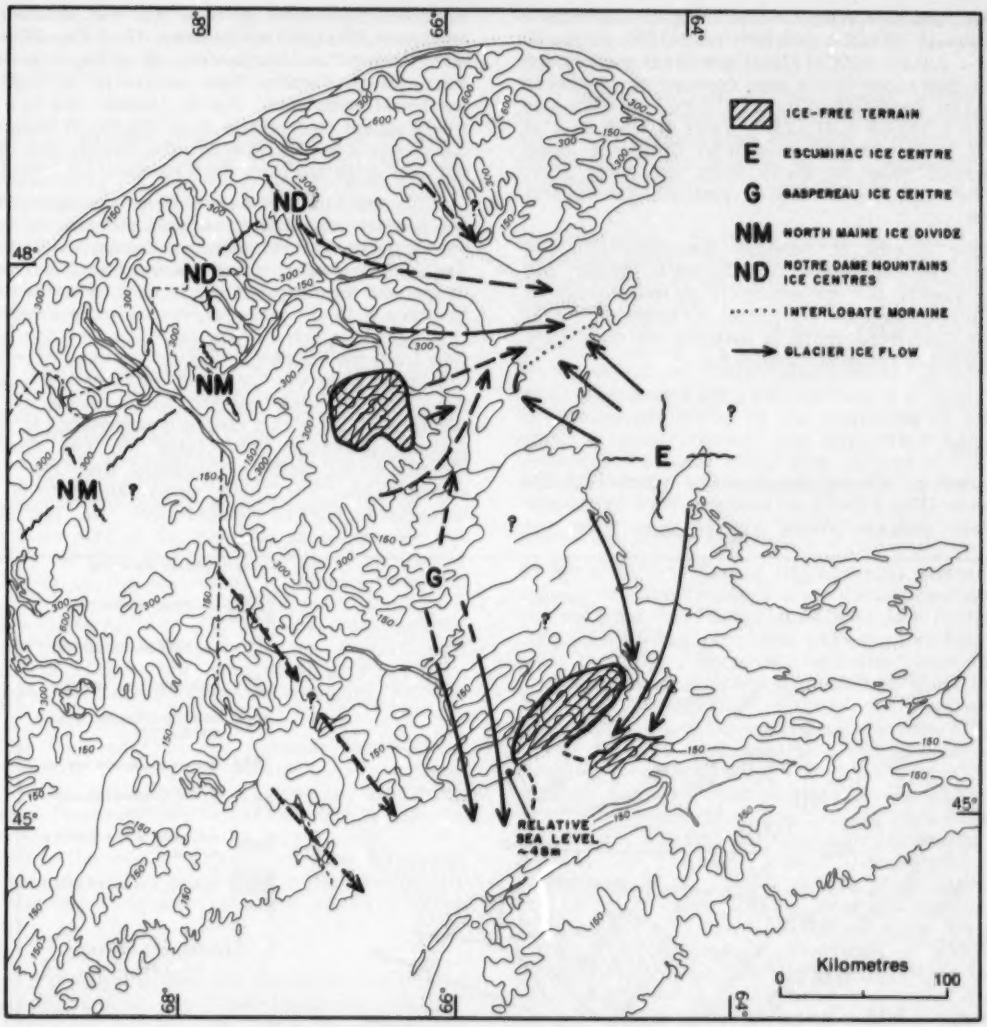


Figure 29. *Paleogeography during Chignecto phase.*

Deltas continued to form adjacent to the ice margin retreating up Chignecto Bay. Glacier retreat was apparently slow and isostatic uplift exceeded any rise in sea level at this time, as the elevations of the deltas decrease to the northeast up the bay. Stea (1983) noted this same phenomenon in adjacent Nova Scotia and also attributed it to slow glacial recession. A similar decline in the elevation of marine features towards Saint John may indicate slow retreat of ice from the low southwest tip of the Caledonian Highlands. It is not clear, however, whether these marine features, e.g. the delta (?) at Black River (Fig. 18) are deltas formed during the retreat of the Chignecto phase glacier or marine benches formed subsequently.

During retreat of the Chignecto phase glaciers from the northern edges of the Caledonian Highlands, short-lived lakes formed in northerly and westerly draining valleys as evidenced by sand-veneered surfaces in these valleys (Rampton and Paradis, 1981c). Accumulations of outwash and alluvium along some of these streams indicate that streams were locally graded to these lake levels or to the glacier margin.

Post-Chignecto Submergence

Sea level rose following retreat of the Chignecto phase glaciers across the Caledonian Highlands northwest of St. Martins and cessation of construction of glacially fed deltas along the Bay of Fundy. Beaches formed at 70 m elevation at Coleraine and at 75 m near Breau Creek (Fig. 18). Small nonglacial deltas formed along Chemical Road and marine silt and clay were deposited to near 50 m elevation north of Riverside-Albert. These features define a maximum level of submergence for the area adjacent to the Bay of Fundy. Isobases constructed from other later sea level indicators to the west (Fig. 30) in southern New Brunswick and Maine, indicate lower levels of submergence in this area at later times.

Chronology

Rampton and Paradis (1981c) had previously dated the Chignecto phase as occurring before 18 ka on the basis of a radiocarbon date on a whale bone (18 570 ± 500 BP; S-1969(A)) in marine clay postdating the Chignecto phase.

Additional dates, however, have lowered the estimated age of this bone to around 12 600 ± 140 BP; (GSC-3284, Table 3). A date 13 600 ± 220 BP (GSC-1222; Table 2) on a mastodon bone in a karst depression in the area covered by Chignecto phase ice suggests deglaciation by that time. A shell was dated at 14 400 ± 530 BP (GSC-2573, Table 2) from east of Saint John (Fig. 21) on terrain covered by Chignecto phase ice, but a second date on shells from this same level (12 800 ± 120 BP, GSC-2640, Table 2) questions the validity of the first date.

The unique history of relative sea level (Table 4) suggests a definite separation of Chignecto phase and Bantalor phase events, but the similarity in soil profiles on drift of each phase suggests a minimal difference of age. Tentatively the Chignecto phase is assigned an early Late Wisconsinan age in the range of 15 to 18 ka.

Bantalor Phase

Ice Flow Patterns

Southwestern Sector. During the Bantalor phase (Fig. 31) striation patterns (Map 1594A) in southern New Brunswick near Saint John indicate strong glacial flow from the northwest (Tay flow pattern, 9 of Fig. 27), from the north-northwest (Norton flow pattern, 12 of Fig. 27), and from the north (Oromocto flow pattern, 20 of Fig. 27). Farther to the west, the St. Stephen flow pattern (21 of Fig. 27) from the northwest dominated. The St. Stephen and Tay flow patterns would appear to originate from the North Maine-Notre Dame Ice Divide complex whereas the Norton and Oromocto flow patterns originate from the Gaspereau Ice Centre.

In the Saint John area, striae related to three patterns are present and suggest variable strength in the ice masses producing these flow patterns. Generally, it appears that the Tay ice flow was strong enough to divert ice from the Gaspereau Ice Centre slightly east of south to produce the Norton ice flow. However, during some interval within the time during which the Bantalor ice was near its maximum, flow from the Gaspereau Ice Centre must have been stronger than flow from the northwest as witnessed by dominance of the Oromocto flow pattern north of Maces Bay and the presence of till with a northern provenance in the same area. Glacial dispersion of clasts in the Clarendon Hills (Thomas et al., in press) suggests that the Tay flow was stronger than the other flows during final deglaciation. Immediately east of Passamaquoddy Bay late glacial flow to

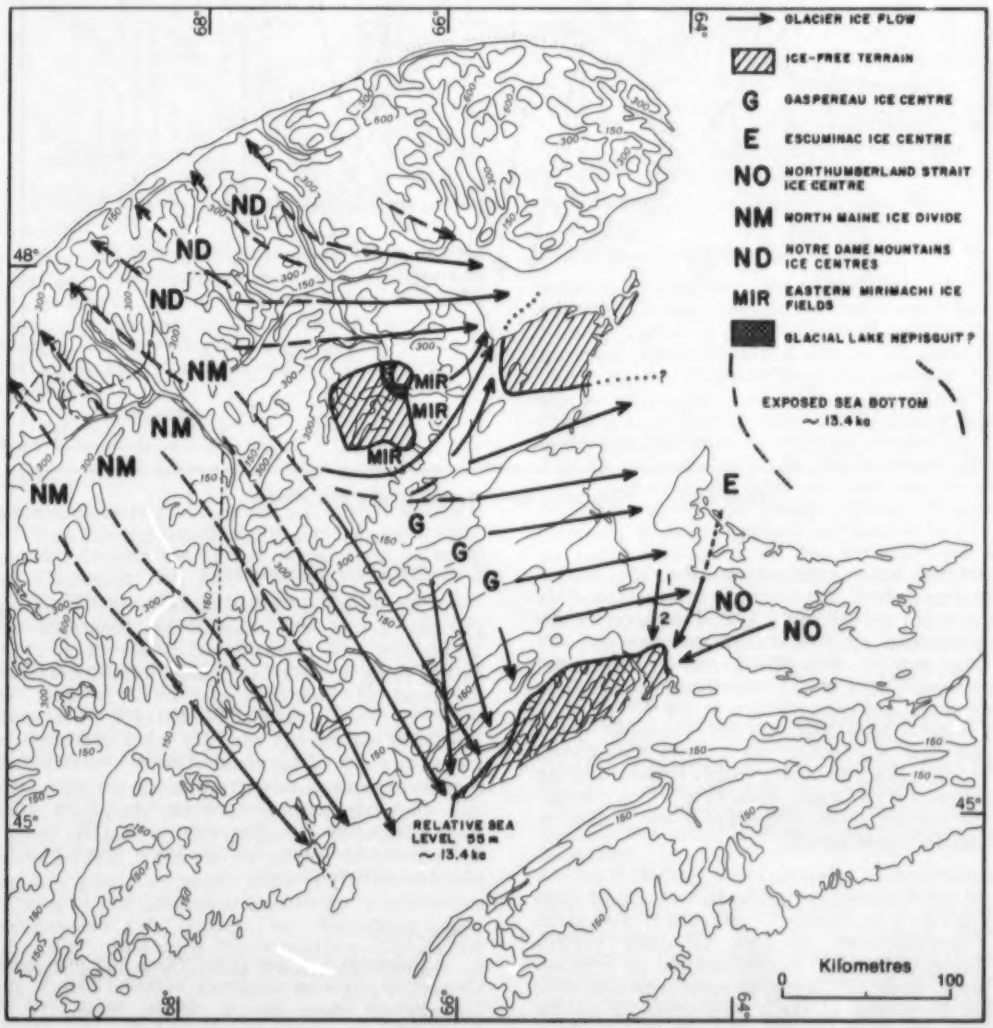


Figure 31. *Paleogeography during maximum extent of Bantalor phase ice.*

the east supersedes the general southeast flow. This diversion (Passamaquoddy diversion, 22 of Fig. 27) is common along the eastern edge of ice retreating towards the North Maine Ice Divide.

Eastern Sector. In eastern New Brunswick the Bantalor maximum was characterized by the strong east-northeast Kent flow pattern (8 of Fig. 27) which had previously been assigned to the Kent phase (Rampton and Paradis, 1981b). The Kent phase has been incorporated into the Bantalor phase because the Bantalor phase is defined to include events occurring while the Gaspereau Ice Centre (source of the Kent flow pattern) was the source of most ice in the New Brunswick Lowlands.

The Kent flow pattern originated primarily from the Gaspereau Ice Centre, although some flow along lower Miramichi River valley may have originated from the North Maine Ice Divide (Fig. 31). The sparse number of igneous erratics to the south of Miramichi River indicates that the area south of the river was relatively unaffected by the ice flowing across the Miramichi Highlands that would have incorporated igneous rocks (Rampton and Paradis, 1981b).

During a later part of the Bantalor phase (termed the Sackville phase by Rampton and Paradis, 1981b, c) ice from a small remnant part of the Escuminac Ice Centre near the north end of Prince Edward Island flowed southward (Tantramar flow pattern, 11 of Fig. 27); ice from another remnant in Northumberland Strait to the east flowed southwestward (Port Elgin flow pattern, 10 of Fig. 27) to form the Memramcook East and Sackville moraines (5 and 6, Fig. 20).

During the final part of the Bantalor phase along Tormentine Peninsula (termed the Tormentine phase by Rampton and Paradis, 1981b) ice flowed generally southward onto Tormentine Peninsula, but encountered ice flowing northward out of Baie Verte at the eastern end of the peninsula (Fig. 27). The relative strength of these two small remnant ice caps wavered as evidenced by the complicated crosscutting relationship between striation patterns there (Map 1594A).

Northern Sector. During the Bantalor maximum, ice flowed to the north-northeast down the Curventon-Bathurst Valley (Nepisiguit flow pattern, 6 of Fig. 27). Along the western and northwestern edges of this lobe, ice flowed from small ice fields on the eastern flank of the Northern Miramichi Highlands (Sevogle and California flow patterns, 16 and 17 of Fig. 27) and, eastward across the Jacquet Plateau (Jacquet flow pattern, 18 of Fig. 27) and down Baie des Chaleurs (Baie des Chaleurs flow pattern, 25 of Fig. 27). Near Bathurst the Jacquet flow pattern was diverted northward by ice in the Curventon-Bathurst Valley. Generally the Nepisiguit flow appears to have been stronger than flow from the west during the Bantalor phase in the Bathurst area. Near Bathurst till having an east-west fabric and containing clasts of the Tetagouche Group, which outcrops to the west, is also overlain by a till with a northeast-southwest fabric and containing clasts of granitic rock, which outcrops to the south. Both tills were probably deposited during the Bantalor phase, although the lower till may have been deposited during the earlier Chignecto phase.

The Jacquet and Baie des Chaleurs flow patterns originated from the North Maine-Notre Dame Ice Divide complex; the Baie des Chaleurs flow was probably fed by ice mainly from the Notre Dame Mountains to the north and northwest (Fig. 31). Most till fabrics along the north side of Baie des Chaleurs indicate an ice source to the north and northwest in that area (Lebuis and David, 1977). The Nepisiguit flow pattern originated mainly from the Gaspereau Ice Centre with some contribution from the North Maine Ice Divide to the west via the Tuadook-Renous ice flow pattern (15 of Fig. 27).

During retreat of the glacier from its Bantalor maximum, ice from the Eastern Miramichi Ice Fields flowed across the western edge of the area that had been covered previously by ice related to the Nepisiguit flow pattern (Gauthier, 1983).

Central Sector. Following retreat of Bantalor ice from its maximum extent, the centre of the Gaspereau Ice Centre shifted slightly to the south. Ice flow was then to the north-northeast (Harcourt and Doaktown flow patterns, 13 and 14 of Fig. 27) over terrain formerly covered by ice flowing to the east. During the final part of the Bantalor phase, when the Gaspereau Ice Centre was breaking up into a number of small independent ice caps, ice flow was partly governed by topography and partly by the positioning of these small ice caps. A small remnant cap near Grand Lake resulted in southwest flow near Minto (Fig. 27) and in the formation of drumlins on morainic topography to the north of Minto.

Glacial and Marine Events

Southern Sector. During its maximum extent, Bantalor ice appears to have formed a discontinuous morainal ridge that extends from Red Head east of Saint John to the south end of Loch Lomond (Latimer Lake moraine, 9 of Fig. 20). East of Loch Lomond ice-contact deposits were deposited by meltwater being diverted down Mispec River (Melvin, 1966). Along the northwestern edge of the Central Plateau the following was noted (Rampton and Paradis, 1981c; p. 26, 28): "Other features associated with the Bantalor maximum are morainal topography along hill slopes east of Hammond, an ablation complex near Hanford Brook, ribbed and rolling moraine between Upham and Cassidy, and a major complex of outwash and ablation drift along the southeastern edge of the Kennebecasis River valley northeast of Sussex (Brinsmead, 1977). It is possible that some ice-contact gravels at the latter locality were deposited as alluvium on the glacier ice by the Kennebecasis River flowing from the glacier-free Caledonian Highlands. Deltaic deposits (Brinsmead, 1977) in this area also may have been deposited by streams flowing from glacier-free areas into lakes along the edge of the glacier. During the maximum extent of the Bantalor phase, defined here as the Hammond River maximum, meltwater probably moved into the upper reaches of Hammond River through gaps in the bedrock ridges east of Cassidy Lake, and at Hillsdale. Drainage from this area may have reoccupied the glacial spillway at Barnsville, but subglacial drainage along the present westward course of the Hammond River is more probable." The Hammond River maximum includes the Hammond River and Sussex East moraines (10 and 11 of Fig. 20). Much of the ablation drift and ice-contact deposits in Little River valley east of Saint John (Melvin, 1966) appears to have been deposited when the ice had withdrawn slightly from its maximum extent.

East of Saint John kame deltas, which probably formed in a glaciomarine environment, are present at elevations of 55 m near the Bantalor ice limit (Fig. 18). Most authors indicate that maximum marine submergence here is at 69 m elevation (Fig. 24); this indicates that sea level was rising more rapidly than the rate of postglacial uplift during the initial retreat of ice in this area (cf. Table 4 and Fig. 30).

Brinsmead and Finamore (1977) noted that the topography of the Anagance Ridges would have isolated blocks of ice during glacial retreat and that late glacial flow may have been affected by the incursion of marine or ponded water. Nevertheless, a number of ice marginal positions,

trending in a northeast direction, can be defined within the Anagance Ridges (Rampton and Paradis, 1981c) on the basis of the position of ablation moraines and ice-contact complexes (Fig. 20).

The topography, persistence of different ice masses (ice from northwest sources appears to have dominated flow during final deglaciation here), and relative elevation of ice-contact deltas indicate that the Nerepis Highlands and Clarendon Hills were deglaciated before the Bocabec Hills, directly to the west, and the western Musquash Lowland (Fig. 20). An ice-contact delta at Browns Flat grades to an elevation of 58 m (indicating a fall in sea level following the marine submergence of the Saint John area to 69 m), whereas a delta constructed in front of ice confined to a valley north of Blagdon grades to an elevation of 65 m (Fig. 18). Further retreat of ice in this valley resulted in a delta being constructed at an elevation of 58 m at Wirral. A similar rise and fall in sea level occurred near Pennfield (marked by t_8 through t_{10} on Fig. 30) where a delta north of Pocologan grades to 53 m, one at Woodland grades to 68 m, at Pennfield Plain to 73.5 m, at Utopia to 66 m, and at McDougall Lake to 62 m (Fig. 18). Maximum submergence of the area, when ice margins were located at Blagdon and Pennfield Plain, indicates a correlation of the ice margins – both marked by significant deltas and associated ice marginal features.

Retreat of the ice margin towards the Gaspereau Ice Centre (north and east) shortly preceded retreat of the ice margin toward the northwest (towards the North Maine Ice Divide) during most of the deglaciation of the Magaguadavic Highlands and Rusagonis Hills. This is suggested by both the Passamaquoddy diversion (22 of Fig. 27) and the bending to the east of the lower segments of some southeasterly trending esker systems; this would only occur if hydrostatic gradients (basically the slope of the glacier surface) changed from a southeasterly trending gradient to an easterly trending gradient near the glacier terminus.

During deglaciation of the St. Croix Highlands numerous ribbed or ridged moraines and large eskers were constructed (Fig. 13, 16). Along St. Croix River valley the drift is thick and generally drumlinized. This suggests that a glacial readvance or surge may have drumlinized former moraines there and that the ice marginal positions depicted in Figure 20 should show more pronounced southerly lobations down St. Croix valley.

During retreat of the ice margin, lakes formed along the northern edge of highlands (Thibault, 1981; Thomas et al., in press). One appears to have formed in the upper part of southern Oromocto River valley (glacial Lake South Oromocto, Fig. 22), although part of the submergence might be due to marine water overtopping the pass at the south end of the valley. The earliest phase of glacial Lake South Oromocto drained through a spillway east of modern South Oromocto Lake (Thomas et al., in press). Further retreat of ice towards the Gaspereau Ice Centre opened drainage to the east. The most significant is the Shin Creek spillway, which drained the area southeast of modern Oromocto Lake (Gadd, 1973).

Lower areas along the coast and along Saint John River valley were submerged by marine waters during glacial retreat. Incursion of marine water up major river valleys may have been more extensive than that shown on Map 1594A, as massive blue clay has been reported to underlie sand along Magaguadavic River valley near Brockway (C. Vail, personal communication, 1982). Alternatively, this clay could be glaciolacustrine – resulting from a dam on the lower Magaguadavic caused by the Passamaquoddy diversion. Subsequent emergence proceeded relatively rapidly (Fig. 24).

Eastern Sector. Few moraines and eskers were constructed in the eastern sector during the early part of the Bantalor phase in areas affected by the Kent flow pattern (8 of Fig. 27), the major exception being the large esker near Blackville (Rampton and Paradis, 1981b). The ice margin along its eastern edge may have been calving into water over most of the area, although Rampton and Paradis (1981b) believed that materials interpreted as outwash and deltaic deposits underlying marine deposits at low elevations along lower Miramichi River may indicate deglaciation prior to submergence.

Figure 32, which shows relative sea level curves for the Gulf of St. Lawrence just east of Prince Edward Island and for Chatham, suggests the possibility that a large freshwater body could have occupied much of the New Brunswick Lowlands during their deglaciation. Kranck (1967) identified submarine benches to depths of 80 m east of Prince Edward Island that she believed were wave-cut prior to 12.7 ka. Rampton and Paradis (1981b) believed that much of the New Brunswick Lowlands were submerged to depths of over 100 m, probably prior to 12.7 ka. Hydrographic charts indicate that a continuous submarine ridge at a depth of about 73 m stretches from Gaspésie to Îles-de-la-Madeleine (Magdalen Islands) and that shallower water exists between Îles-de-la-Madeleine and Nova Scotia (Fig. 33). This topographic

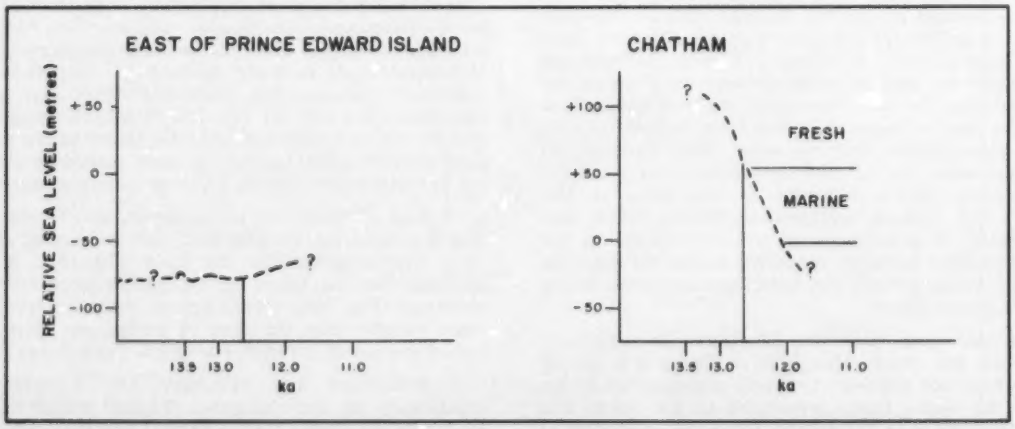


Figure 32. *Deduced relative sea level curves for east of Prince Edward Island and for Chatham, New Brunswick. Curves based on data from Kranck (1967) and Rampton and Paradis (1981b). Assume general eustatic depression in sea level of about 80 m around 13 ka.*

configuration, combined with isostatic downwarping, might have produced a lake in the New Brunswick Lowlands during their deglaciation before 12.7 ka. In any case, isostatic depression of the glaciated New Brunswick Lowlands, the general low slope of this land mass, and a low relative sea level would result in a shallow sea with low salinity and possibly a depauparate biota in the area immediately following deglaciation.

On the Chignecto Isthmus a major ribbed moraine was constructed east of Memramcook and a hummocky moraine at Sackville (6 and 5 of Fig. 20) by ice flowing from remnants of the once larger Escuminac Ice Centre. The moraine at Sackville is composed largely of sandy material and may have formed in a subaqueous environment. Finally, the Tormentine moraine (28 of Fig. 20) was constructed by ice impinging only on Tormentine Peninsula (Rampton and Paradis, 1981b).

Northern Sector. During the Bantalor phase, glacial ice is presumed to have reached down the Curventon-Bathurst Valley to Baie des Chaleurs. Along its western flank, Bantalor ice interacted with ice flowing across the Jacquet Plateau and down Baie des Chaleurs; along its eastern edge, it possibly formed the Bass Lake moraine (3 of Fig. 20) at its maximum extent (Gauthier and Cormier, 1977; Finamore, 1978). Eventual retreat of the easterly flowing and north-northwesterly flowing ice masses appears to have resulted in the large ice-contact interlobate mass at Blue Mountain (4 of Fig. 20).

As ice retreated westward across the Jacquet Plateau, proglacial outwash and deltas probably formed at sea level. Former sea level is now marked by extensive gravel deposits west of Nepisiguit Bay (Fig. 18). Generally, no good foresets are present in these deposits, but Finamore (1978) implied the presence of glaciofluvial deposits under marine gravels in this area. This suggests a rapidly rising sea level concurrent with deglaciation (Fig. 30). Ablation moraine at Nigadoo Lake and ice-contact deposits to the south may have formed during temporary stillstands in the retreat of the ice margin from the area (Finamore, 1978).

Farther to the north, a suture appears to have developed in the zone of contact between the Baie des Chaleurs and Jacquet flow patterns (25 and 18 of Fig. 27). The Elmtree interlobate moraine (27 of Fig. 20), comprising

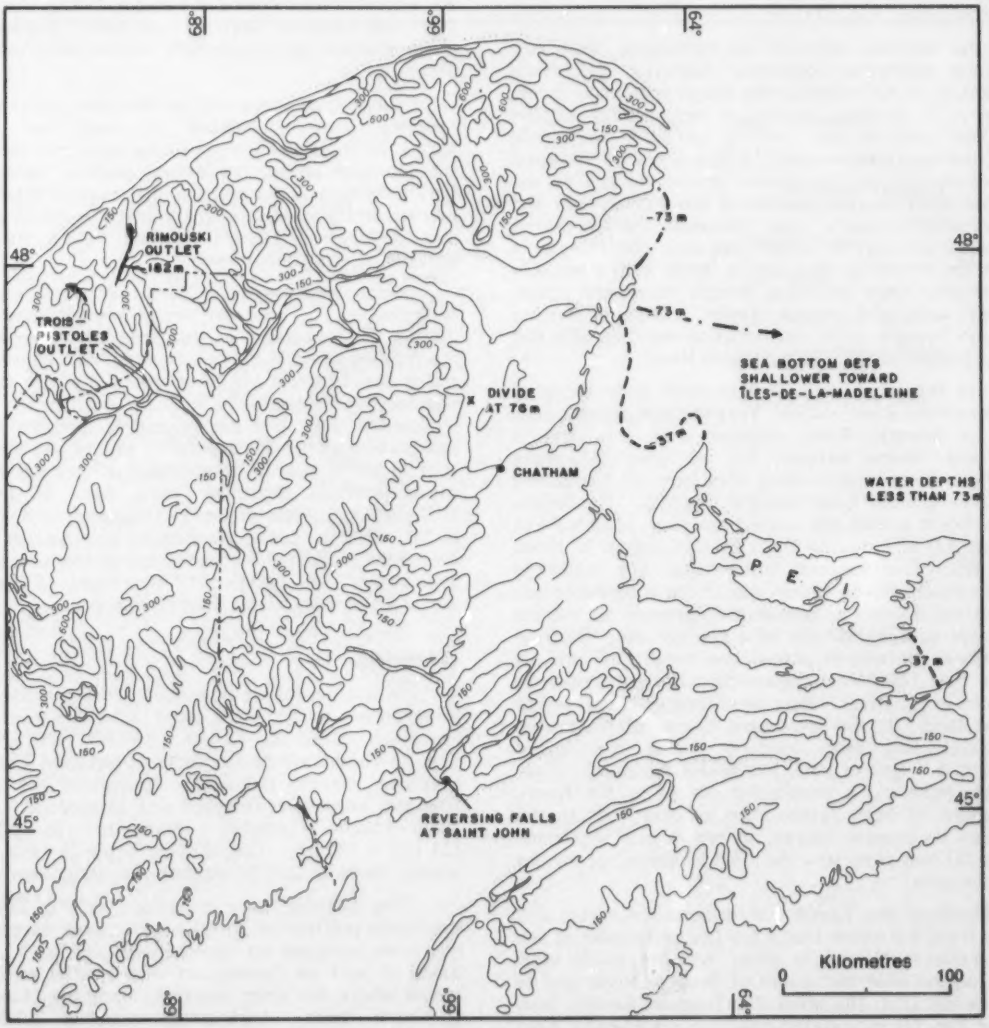


Figure 33. *Location and elevation of significant submarine and terrestrial features affecting Quaternary water bodies in New Brunswick and adjacent areas.*

ablation till, ice-contact deposits and outwash, and possibly glaciomarine deposits (Gauthier and Cormier, 1977), forms a major interlobate feature marking meltwater drainage to the east between these two lobes of ice. Clasts within this feature have both inland and coastal provenances, indicating a component of ice flow from both directions during its construction. Some late southward trending striae (Map 1594A) and a northward concavity of the western segments of this ridge suggest that the Baie des Chaleurs ice was the last ice mass to retreat from parts of this moraine.

During the southward retreat of the Nepisiguit lobe (6 of Fig. 27), it appears to have been calving into water because much of the area is marked by DeGeer moraine (Map 1594A), which is characteristic of terrain where glaciers have terminated in water (Prest et al., 1968). In the Curventon-Bathurst Valley the ice probably terminated in marine waters, although shell dates from adjacent areas on the Acadian Peninsula and Central Lowland (Fig. 19) all postdate 12.7 ka, and suggest that saline waters may not have transgressed this area until after this time. To the east, the ice front appears to have terminated in lakes confined between the glacier and higher land to the north, or between the Nepisiguit lobe and remnant ice along Miramichi Bay (glacial Lake Bartibog, Fig. 22).

Along the western edge of the retreating Nepisiguit lobe, numerous meltwater channels, morainal and esker ridges, in addition to the crosscutting striae previously cited, are evidence of a complex pattern of deglaciation of the Nepisiguit lobe and of ice moving off the Miramichi Highlands to the west. Brinsmead (1978) has traced morainal systems representing possible pauses in the retreat of the Nepisiguit lobe south of the junction of the Tomogonops and Northwest Miramichi rivers: one intersects the Northwest Miramichi valley axis at the above junction; one intersects the valley at the mouth of McLaughlin Brook with a possible equivalent morainal ridge damming Sevogle River; one passes through Estey Lake and possibly under marine sediments adjacent to the Sevogle delta (for location see Fig. 18); and finally, one is present along Little Sevogle River.

As the ice retreated southward, a water body occupied Northwest Miramichi River valley. Varved clays present near the mouth of Sevogle River suggest that this was a freshwater lake which existed for at least 370 years (370 varves have been noted) along this part of Northwest Miramichi River (glacial Lake Sevogle, Fig. 22). The height of land at the north end of this valley, however, is only 76 m elevation (Fig. 33) and marine limit in this region is about 120 m (Fig. 30). This suggests either that the water in Northwest Miramichi River valley was at the freshwater end of a long estuary or that sea level transgressed to marine limit subsequent to development of a glacial lake. Because no major spillway developed across the height of land, it would appear that Northwest Miramichi valley was open to the sea via Miramichi River valley when sea level fell below the height of land. If the southward course of Northwest Miramichi River had been blocked during the initial emergence of this height of land, Northwest Miramichi River would probably have cut a permanent course to the north, across the height of land, rather than bending 180° to the south to occupy its present course. A brown clay with sand partings (Fig. 21) may represent the final incursion of marine waters into this area.

Final decay of the Eastern Miramichi Ice Fields and retreat of ice from the North Maine Ice Divide to west of the mouth of Northwest Miramichi River was relatively late. Deltas were formed near the mouth of Sevogle River and at Devils Back Brook (Fig. 18) when the Tuadook-Renous flow pattern (15 of Fig. 27) terminated near the Breadabane West moraine (26 of Fig. 20); these indicate a rapidly falling relative sea level at this time (i.e., postglacial uplift proceeding more rapidly than rising sea level, Fig. 30). This is characteristic of the latter part of the Bantalor phase.

In northwestern Maine glacier flow was probably to the north-northwest or northwest via the Lac Baker ice flow pattern (24 of Fig. 27) from the North Maine Ice Divide as proposed by Kite et al. (1982). Most certainly, parts of lower St. Lawrence valley were deglaciated by this time (Locat, 1977) and the effect of Laurentide ice in northern New Brunswick was negligible.

Central Sector. During the latter part of the Bantalor phase the Gaspereau Ice Centre changed to an ice divide with many domes connected by sags. Glacial striation patterns between Newcastle and Harcourt (Map 1594A) suggest that ice moving from the west interacted with ice moving from the south. Later the centre of Gaspereau ice flow shifted farther west (Rampton and Paradis, 1981a, b). Pauses in glacial retreat or short-lived re-advances may have resulted in construction of the New Canaan and Sunbury moraines (8 and 25 of Fig. 20). Along the northern flank glacier flow and hydrostatic gradients shifted to the north and northward-trending eskers were formed (Rampton and Paradis, 1981b) in areas where ice flow had formerly been to the east. During most of this phase glaciers were generally calving into water, marine or otherwise.

Finally, as suggested by striation patterns, glacial flow radiated from a number of small ice caps (Fig. 20, Map 1594A). These ice caps may have persisted to the end of the Bantalor phase, preventing marine inundation of local low-lying areas. Rampton and Paradis (1981a) believed that marine limits are discordantly lower in the Grand Lake Basin and southwestern Central Lowland than in surrounding areas because of this phenomenon.

Near Napadogan, deglaciation was complex: "Striae and fluting observed on aerial photographs indicate two distinct flow patterns in this area .... It appears that when ice thinned and the Renous flow pattern ceased to exist, ice either moved south off the edge of the high ground north of the Southwest Miramichi River or north from the Gaspereau ice centre. The late ice movement temporarily blocked the headwaters of the Southwest Miramichi and allowed a glacial lake to form west of Napadogan .... during the last part of the Gaspereau phase or early part of the subsequent Millville/Dungarvon phase." (Rampton and Paradis, 1981a, p. 23). Glacial Lakes Napadogan and Juniper (Fig. 22) may owe their existence to depression of terrain occupied by the headwaters of Southwest Miramichi River below the elevation of a sill on its downstream course.

Chronology

In southern New Brunswick, fossiliferous till in a ridge immediately south of the Saint John moraine (14 of Fig. 20) has been dated at 13 100 ± 160 BP (GSC-3557, Table 2). This, in combination with dates of between 13 and 13.2 ka on overlying marine sediments throughout the area, suggests that the Saint John moraine was formed ca. 13.2 ka. Dates from basal organic sediments in Cassidy Lake (13 100 ± 280 BP, GSC-2787; Table 2) and palynological studies by Mott (1975b) support this assignment.

The Latimer Lake moraine (9 of Fig. 20), marking the maximum position of the Bantalor phase east of Saint John, has been assigned an approximate age of 13.4 ka. This is based in part on correlation with glacial events in coastal Maine where the older moraines along the coast are believed to have been constructed between 13.5 and 13.3 ka (Borns, 1973; Thompson, 1982). The maximum extent of the Bantalor phase as shown in Figure 31 has been assigned a similar age.

The ice margin defined by the Pennfield and Blagdon deltas has been assigned an age of 13 ka, based on the oldest date on shells in sediments overlying associated glaciomarine deposits. Perhaps the ice margin would have left this area shortly prior to that time.

Scott and Medioli (1980) have suggested that much of the New Brunswick coast west of Saint John was submerged by marine water before 16 ka. This is based on total carbon dates from cores, which show the transition between foraminiferal assemblages indicating marine conditions and those showing brackish conditions in lake basins. The senior author feels that these samples must have been contaminated by older carbon. Mott (1975b) has demonstrated through palynological methods that radiocarbon dates (GSC-1063, -1272, -1447, -1604; Fig. 19, Table 2) obtained from a lake near Pennfield (Gadd, 1973) were similarly in error by as much as 3000 radiocarbon years. In addition, Amos et al. (1980) obtained a date of 2200 ± 110 BP on modern material deposited in a modern salt marsh at the Bay of Fundy using the total carbon method; Fillon et al. (1981) documented a discrepancy of more than 4000 radiocarbon years between dates on shells and on total organic material.

Most of the New Brunswick Lowlands are believed to have been deglaciated by 13 ka as they appear to have been submerged to over 100 m elevation (Fig. 18). This high level of submergence requires that the area be deglaciated before rapid postglacial uplift began.

Millville/Dungarvon Phase

Ice Flow Patterns

During the Millville/Dungarvon phase, ice on the Chaleur Uplands adjacent to Saint John River valley had a southward slope as is reflected by the directional trend of eskers and the Woodstock flow pattern (23 of Fig. 27). Earlier flow had probably been in a more southeasterly direction as suggested by the striation summary (Map 1594A) and Figure 27. Thompson (1982) noted this same chronological sequence of ice flow directions in adjacent southern Maine. Only at the very northern end of the Central Miramichi Highlands did glacier ice spill over to the east (Tuadook-Renous flow pattern, 15 of Fig. 27).

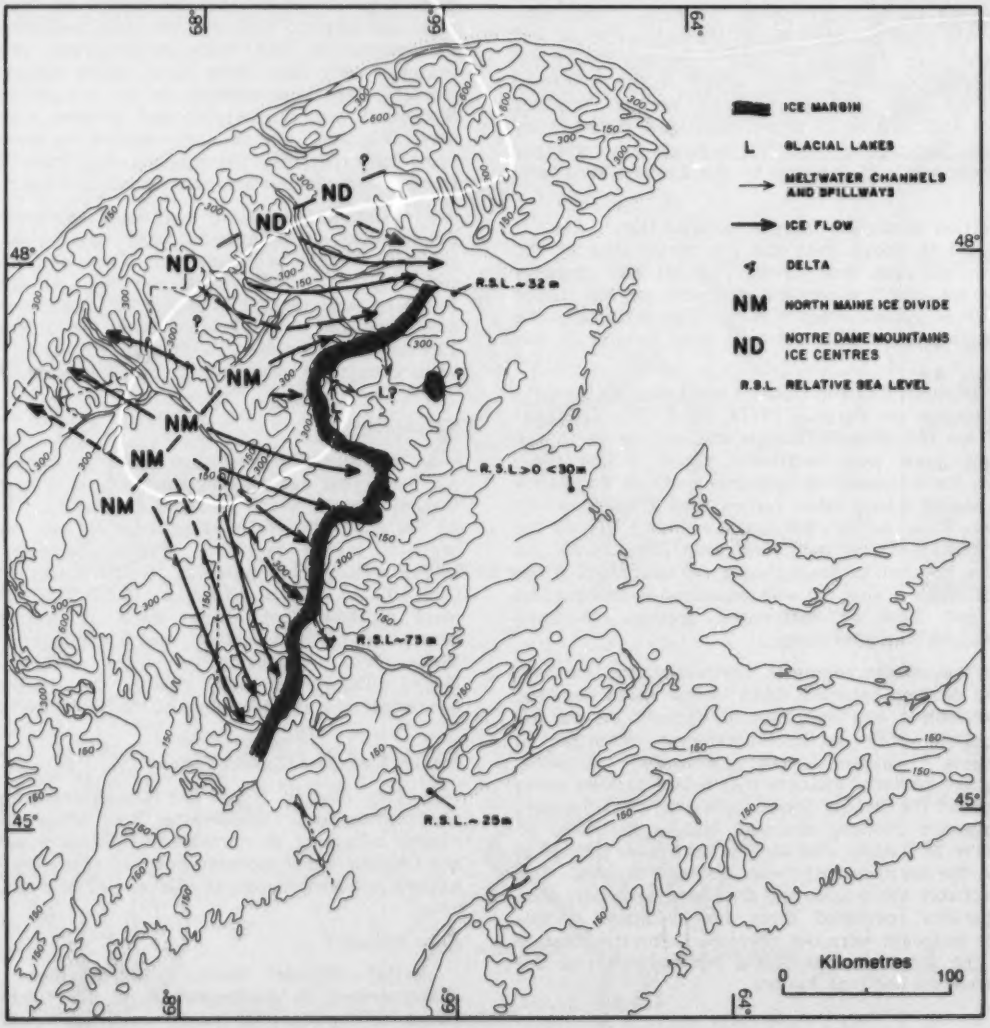


Figure 34. *Paleogeography during maximum extent of Millville/Dungarvon phase ice.*

North and west of the Northern Miramichi Highlands ice flow patterns probably continued to follow those established in the earlier Bantalor phase, namely eastward towards Baie des Chaleurs and northwestward to St. Lawrence River (Fig. 34).

Glacial and Marine Events

Glacial Margins. Major ribbed moraine along a belt marking the maximum extent of glacier ice during the Millville/Dungarvon phase and associated extensive valley trains and glaciofluvial deltas indicate that the glaciers may have paused for a significant period of time at this ice margin. The McAdam, Millville, Dungarvon, and Serpentine Lake morainal belts (29, 30, 31, 32 of Fig. 20) and Zealand delta (Fig. 18) are associated with this maximum stand. The ribbed moraine (Fig. 14) backed by fluted and striated topography indicates the presence of sporadically active ice during parts of this phase (Rampton and Paradis, 1981a), even though eskers suggest a stagnating, downwasting ice sheet (Finamore, 1981).

During the maximum extent of Millville/Dungarvon ice (Fig. 34), meltwater drained down Serpentine River to near Serpentine Mountain where a lake, glacial Lake Mamozekel (Fig. 22), existed for a short period of time. This lake drained over the pass between South Mamozekel River and Nictau Lake to glacial Lake Nictau (Fig. 22; Brinsmead, 1979a; Gauthier, 1983), which in turn drained eastward into the upper Nepisiguit drainage system. Whether a glacial lake still existed in upper Nepisiguit River valley (glacial Lake Nepisiguit, Fig. 22) depends on whether Nepisiguit River was still dammed by the Eastern Miramichi Ice Fields.

The position of the ice margin north of Nictau Lake is unclear. Figure 34 shows that the ice terminated at the Three Corner morainal belt (34 of Fig. 20) and probably formed some of the ice-contact deposits on the upper Tetagouche River system (Map 1594A). The delta near the mouth of Jacquet River (Fig. 18) may have formed at this time.

On the southern Chaleur Uplands the following scenario developed (Rampton and Paradis, 1981a, p. 25, 27): "Retreat of the ice from the Millville/Dungarvon ice margin to the Tobique River area was relatively rapid. Lee (1962) suggested that the ice paused temporarily north of Woodstock at the south end of a long esker system that discharges into the Saint John River valley, but little evidence of this ice margin is present east of the Saint John River. An ice margin position also can be traced along the east flank of the basin north of Bath. Thick till was deposited in this part of the basin, and flow of meltwater appears to have concentrated along its eastern edge.

"As the ice margin retreated northward, the ice may have begun to develop separate lobes in low areas between hills south of Juniper and northeast of Kilfoil. As the ice margin retreated even further northwest drainage in northwest-sloping valleys was probably dammed. Aerial photograph patterns indicate that a lake formed along the ice margin in the Beadle Brook valley north of Juniper. Meltwater channels crossing drainage divides northeast of Kilfoil and north of Juniper also suggest that lakes formed in the valleys to the northwest of these drainage divides. The lakes were probably short-lived and drained subglacially soon after the glaciers retreated from the drainage divides because major spillways were not developed with the possible exception of the upper reaches of the North Branch of the Southwest Miramichi north of Juniper.

"Kame terraces, kame deltas and other ice-contact deposits were deposited along the margin of retreating ice within the confines of the incised Saint John River valley (cf. Lee, 1962; Finamore, in press). Detailed mapping (Lee, 1962; Finamore, 1981; Thibault, 1980) in the area indicates that most ice-contact deposits within the Saint John River valley are south of Hartland." Kiewiet de Jonge (1951) believed that much of the glaciofluvial material deposited within the confines of Saint John River valley was deposited subaqueously.

Relative Sea Levels. Relative sea level was probably near 32 m at the mouth of Jacquet River and lower than 30 m over the Central Lowland (Fig. 30) when the Millville/Dungarvon ice stood near its maximum. During this interval the submerged parts of northern and eastern New Brunswick were certainly covered by saline waters because shells and bones from these areas and adjacent Prince Edward Island (Prest, 1973) all date from between 12.7 and 12.4 ka (Fig. 19, Table 3); the only exception is a date of 12 110 ± 200 BP (BGS-514; Table 3) on shells east of Bathurst.

Relative sea level stood near 73 m to the west of Fredericton and was well above modern sea level at Saint John (Fig. 24) when Millville/Dungarvon ice was near its maximum extent. Relative sea level undoubtedly fell during the course of the Millville/Dungarvon phase. Deltaic features along Saint John River valley suggest that residual isostatic depression allowed the sea to reach near 82 m when Woodstock was deglaciated and greater elevations to the north (Fig. 18). The deduced relative sea level curve for the Saint John area (Fig. 24) suggests that Saint John valley was flooded with saline water until about 12.3 ka.

Most marine sediments at Saint John and west along the Fundy coast were deposited prior to 12.3 ka. At this time isostatic rebound had resulted in a sea level lowering to near the elevation of the modern coastline. All dated shells from this area predate 12.3 ka (Table 2). In adjacent Maine, shell dates reported by Stuiver and Borns (1975) generally predate 12.2 ka, and Schnitker (1974) suggested that marine retreat to this coastline was complete by 12.3 ka.

Chronology

The Millville/Dungarvon phase has been dated at about 12.7 ka. This age is assigned because it must predate the Chaleur phase, which is dated at 12.4 ka. Also, an age of 12 700 ± 200 BP (BGS-684; Table 2) was obtained on an organic layer overlying glacial deposits inside the Millville/Dungarvon glacial limit south of Woodstock (Fig. 19). A date of 13 800 ± 170 BP (GSC-657; Table 3) on marl southwest of Grand Falls is erroneously old as associated gyttjas dated much younger (Lowdon and Blake, 1970). Most other dates within the limit of Millville/Dungarvon ice relate to marine events or submergence and to drainage of the Saint John River system.

Plaster Rock and Chaleur Phases

The Plaster Rock and Chaleur phases are roughly contemporaneous. The Plaster Rock phase concerns glacial events occurring in northwestern New Brunswick, whereas the Chaleur phase concerns events occurring largely in the eastern and northern areas adjacent to Baie des Chaleurs.

Flow Patterns

The Plaster Rock and Chaleur phases were characterized by disintegration of the North Maine-Notre Dame ice complex into small, shrinking, but active, ice caps

in northwestern New Brunswick. Stratigraphic evidence along Baie des Chaleurs indicates that ice advanced eastward into Baie des Chaleurs (Baie des Chaleurs flow pattern, 25 of Fig. 27), fed in part by ice from Matapedia River valley (Gauthier, 1983). Ice probably also flowed out of Notre Dame Mountains, at least along the major valleys (Gounamitz flow pattern, 26 of Fig. 27), during the early part of the Plaster Rock phase. Meltwater channels and crosscutting striae (Map 1594A) indicate that later ice flowed northward from the North Maine Ice Divide into the southeastern edge of the Edmundston Highlands. Northwestward flow (Lac Baker flow pattern, 24 of Fig. 27) probably persisted for the early part of the Plaster Rock phase near Edmundston.

After the break up of the North Maine Ice Divide, ice probably radiated from the St-Quentin Ice Centre (Fig. 35). Striae indicating flow from the St-Quentin Ice Centre are rare, but moraines, eskers, and meltwater channels (Fig. 36) indicate that ice sloped away from this centre. Small glacier remnants within valleys may have continued to flow during the final part of the Plaster Rock phase. Gauthier (1982) recorded striae with a southeastward trend in the lower part of Green River valley and near Saint-Jacques in Madawaska River valley; Gauthier (1983) believed that these were formed as a result of southeastward flow towards a rapidly calving glacier front in Saint John River valley southeast of Edmundston.

Early Chaleur Phase

Diamicton and glaciofluvial deposits overlie fossiliferous marine clay between Charlo River and Dickie Cove on Baie des Chaleurs. This indicates either that a glacier overrode the clays or that an increasing abundance of dropstones was being deposited in a glaciomarine environment because of a glacier advance into Baie des Chaleurs. Alcock (1936) also reported deformed varved clay under till along Baie des Chaleurs. The presence of ice-contact glaciofluvial deposits over the diamicton suggests that the clays were overridden by glacier ice, even though pebbles within the diamicton at Dickie Cove and Durham Centre (sites of GSC-3485 and GSC-3552, Fig. 19) do not show good till fabrics. Perhaps, the glacier surged into the bay and was in a floating mode at these sites. Before this re-advance or surge, the glacier may have retreated to the west of Campbellton where gravel has been observed beneath till.

This glacier advance dammed or diverted rivers flowing from the Chaleur Uplands (Fig. 35a), causing kames and kame deltas to form along the edge of the Chaleur coastal plain. The mouth of Jacquet River was free of ice at this time; whether it was still receiving meltwater from its headwaters is not clear. A suture developed along the south side of the glacial lobe in Baie des Chaleurs and a large interlobate system developed consisting of ice-contact glaciofluvial sediments and ablation till (Brinsmead, 1979b). Much of the material in this interlobate system, the 'Restigouche Kame' (37 of Fig. 20), was deposited subaqueously or as interlobate

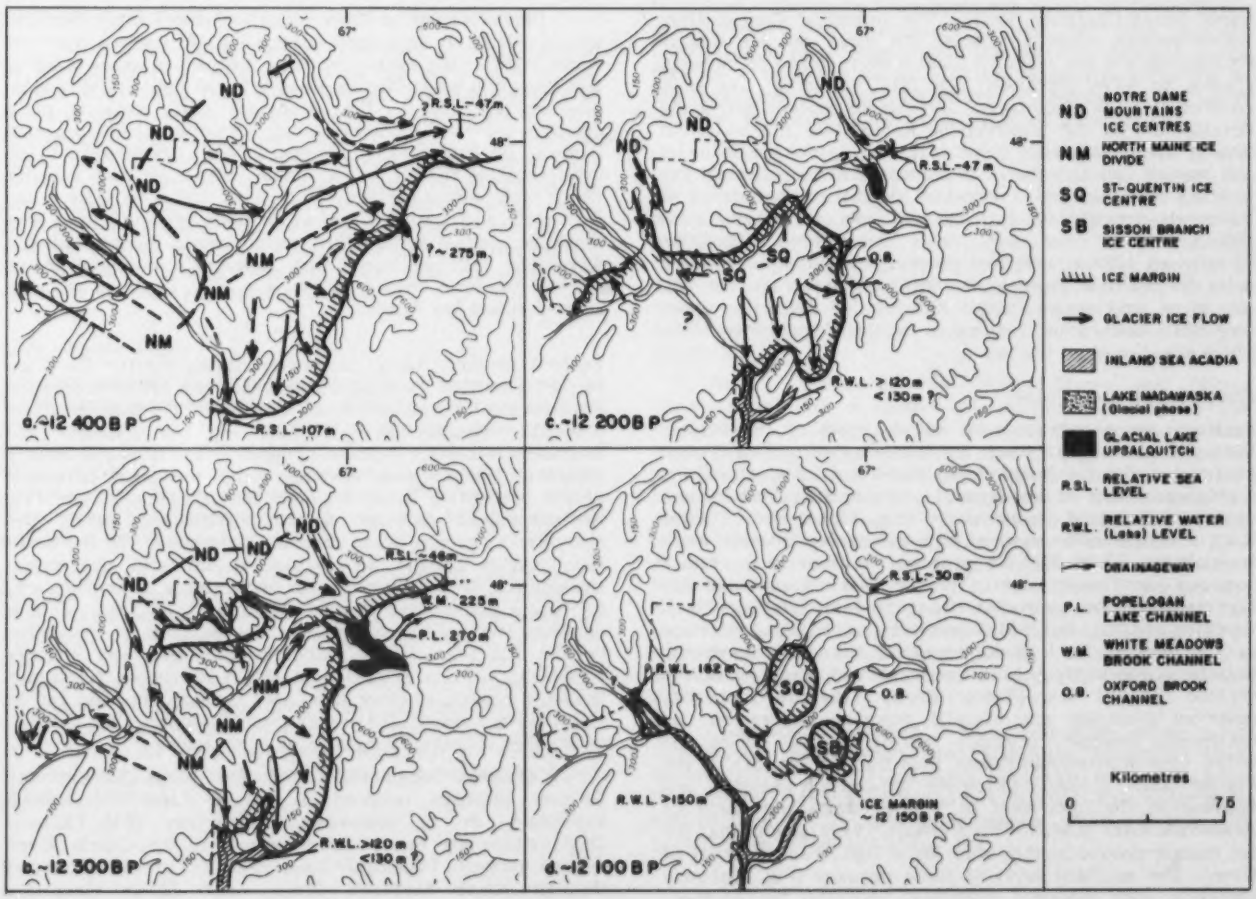


Figure 35. *Paleogeography during Plaster Rock and Chaleur phases.*

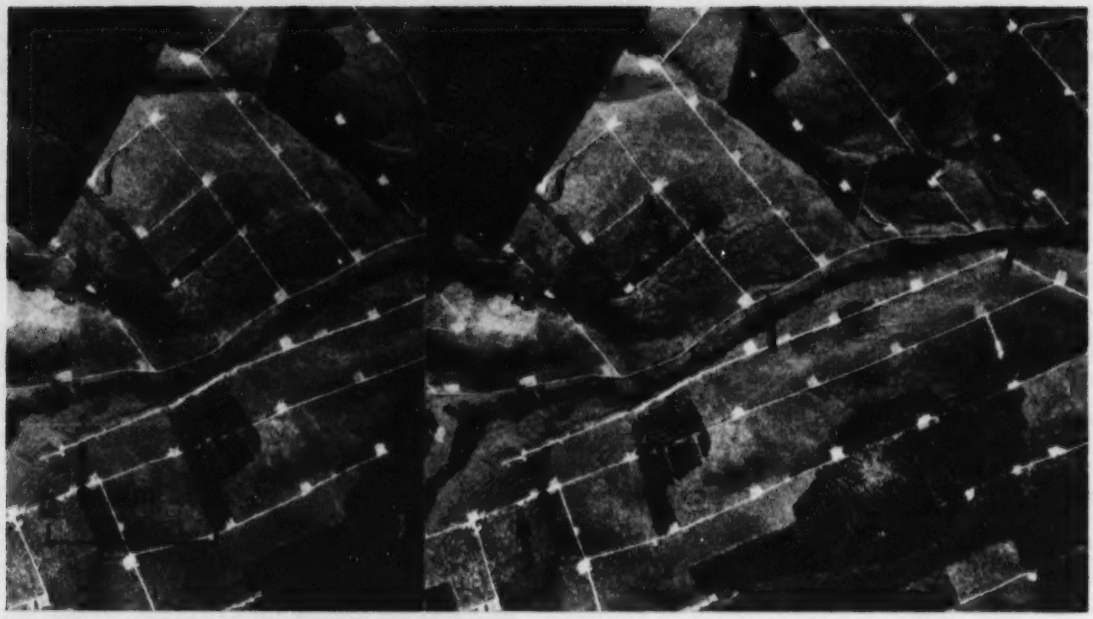


Figure 36. *Meltwater channels on valley walls west of Sisson Branch Reservoir. Channels are obvious only in clear-cut areas (NAPL A25021-92, 93).*

kame deltas (Gauthier, 1983). The mouth of Charlo River valley became clear of glacier ice when the ice margin receded up the bay (Fig. 35b) and a delta began to develop, grading to 47 m elevation (Fig. 18). Ice may have been covered by thick glaciofluvial and marine deposits during the development of the 'Restigouche Kame' near the mouth of Charlo River, as kettles have developed in the glaciofluvial and marine deposits there (Brinsmead, 1979b). During this interval morainic and ice-contact deposits accumulated at Balmoral, deposited mainly by ice from the north, even though they occupy an interlobate position (Brinsmead, 1979a). With further retreat a glaciomarine delta formed near Dundee at about 46 m elevation (Fig. 18), also in an interlobate position between ice occupying upper Baie des Chaleurs and ice draped on the Campbellton Hills, which stand south of the bay.

Early Plaster Rock Phase

Southern Sector. During the earliest part of the Plaster Rock phase, the North Maine Ice Divide appears to have been continuous with the Notre Dame Mountains Ice Centres, and a continuous belt of ice-contact and morainal deposits was constructed around its periphery (Fig. 35a). Near Plaster Rock a unique esker system with tributaries (Plaster Rock interlobate, 35 of Fig. 20) probably formed in a suture between ice flowing down Tobique River valley and ice flowing from the northwest (Rampton and Paradis, 1981a). Rampton and Paradis (1981a) believed that the last active ice in the Plaster Rock area advanced from the northwest because brown sandy till has been observed overlying locally derived red till near Plaster Rock. The till that they observed overlying glaciofluvial deposits farther to the northwest, however, probably represents flow during the entire Late Wisconsinan rather than that during part of the Plaster Rock phase. Extensive ice-contact deposits and outwash in the hills west of Plaster Rock and north of Aroostook River (Lee, 1959b; Thibault, 1980) suggest that the ice margin paused in this area for a significant interval of time. The southern limit of these deposits (Fig. 35a) may correlate with morainal complexes defining the southern limit of the Van Buren Till in adjacent Maine (Genes and Newman, 1979, 1980).

Deltas formed at 106 m elevation near Tobique Narrows and at 107 m on Aroostook River (Fig. 18) suggest that the water level of the estuary in Saint John River valley stood at this elevation when Plaster Rock ice was near its maximum position (Fig. 35a). The Aroostook delta consists of a complex sequence of foresets and horizontally bedded gravel (Kiewiet de Jonge, 1951; Lougee, 1954). Consequently, the interpretation that it was a simple delta built into a major water body with a surface elevation of 107 m must be taken with some reservation. This water body became a long lake (initially containing brackish water) when the site of the Reversing Falls at Saint John apparently emerged above relative sea level around 12.3 ka (Fig. 24). This lake has been termed Inland Sea Acadia.

Eastern Sector. Along its eastern flank, Plaster Rock ice formed a belt of ice marginal deposits in a broad depression between the Gulquac and Black Mountains, and Blue Mountain (Gulquac moraine, 36 of Fig. 20; Fig. 17). Eskers and meltwater channels indicate complex easterly and southerly subglacial and proglacial drainage within this area. Diversion of the Serpentine River drainage through Nictau Lake into the Nepisiguit drainage system established during the Millville/Dungarvon phase probably continued. To the north the location of the ice margin is not well delineated. Upsalquitch River, however, was undoubtedly dammed by ice, forming a glacial lake in its headwaters (Fig. 35a). This is demonstrated by the presence of varved clays in the valley bottom (Gauthier, 1979). Whether this lake was initially continuous with a glacial lake (?) in the upper Nepisiguit drainage system, whether it drained through the wind gap at Upsalquitch Lake at 275 m elevation (Gauthier, 1979, 1980a), or whether it drained subglacially, is not known.

Continued deglaciation of the Saint-Quentin and Jacquet plateaus resulted in glacial Lake Upsalquitch expanding in a downstream direction (Fig. 35a,b,c). Deglaciation also opened lower outlets to the Charlo River system through Popelogan Lake near 270 m elevation and through White Meadows Brook near 225 m elevation. Gauthier (1982) proposed drainage of glacial Lake Upsalquitch to the southwest through the Oxford Brook

channel (Fig. 35c) near 225 m elevation. However, morainic topography in Tobique River valley and evidence of a glacial lake near Sisson Branch Reservoir require that ice blocked Tobique drainage until after most of the terrain to the north was deglaciated. Indeed it is envisaged that after retreat of Plaster Rock ice uncovered the Oxford Brook channel, the upper tributaries of Tobique River drained to the north through this channel until Tobique River valley was completely deglaciated (Fig. 35c).

Northwestern Sector. During the early part of the Plaster Rock and Chaleur phases, ice from the St-Quentin Ice Centre and Notre Dame Mountains Ice Centres probably coalesced and flowed northeastward towards Baie des Chaleurs (Fig. 35a). Ice flow across the Kedgwick Highlands may have been minimal as the eastern parts of the highlands are greatly dissected and are characterized by V-shaped valleys, even though Kedgwick valley itself is characterized by glacially polished and striated bedrock. A poorly developed interlobate moraine with numerous associated meltwater channels along Grog Brook to the northeast of Kedgwick was deposited as the ice margins retreated southward towards Kedgwick and westward up Patapedia River valley (Fig. 35b). Relatively early in the Plaster Rock phase the higher parts of the Kedgwick Highlands may have been deglaciated, with ice from the Notre Dame Ice Centres impinging on the northern part of the high ground and ice from North Maine Ice Divide impinging on their southern edge (Fig. 35b). The limit of northerly flowing ice from the North Maine Ice Divide is marked by late northward trending striae (Striation Summary, Map 1594A) and meltwater channels which carried water north and west along the edge of the ice (Map 1594A).

Most of the Saint-François Hills were probably covered by northwesterly flowing ice during the early part of the Plaster Rock phase (Fig. 35a). Retreat and downwastage of the ice would result in the most western part of New Brunswick being deglaciated first (Fig. 35b). Eskers and paleocurrent directions in glaciofluvial deposits near Lac-Baker indicate an ice margin continuously retreating to the east during deglaciation of the Saint-François Hills (Fig. 35b,c,d). Lakes dammed against this retreating ice probably drained northward through the Cabano River system in Quebec. Similar drainage has been proposed for northwestern Maine (Kite et al., 1982).

Late Chaleur Phase

Ice appears to have persisted for a long interval in lower Restigouche and Matapedia valleys at the head of Baie des Chaleurs (Fig. 35c). Shrinkage of this ice lobe northward to near the mouth of Upsalquitch River would have allowed glacial Lake Upsalquitch to drain initially across a drainage divide near 137 m elevation into Christopher Brook (Finamore, 1979a) and then across one near 105 m elevation into Black Brook (Gauthier, 1983).

Deglaciation of the mouth of Christopher Brook valley resulted in construction of a delta, graded to sea level, at 46 m elevation near here. A delta was constructed at Flatlands, which graded to sea level at 46.5 m elevation (Fig. 18), after ice retreat up Restigouche valley. This suggests that upon final disintegration of the Chaleur lobe, sea water invaded the lower part of Restigouche and Upsalquitch valleys, and that clays observed under alluvium at Robinsonville may be either estuarine or lacustrine. Similar observations have been made in Restigouche River valley by Lebuis (1975).

Late Plaster Rock Phase

The late part of the Plaster Rock phase is characterized by the production of patches of ablation moraine, abundant ice-contact deposits, and meltwater channels.

Southern Sector. As the ice margin retreated north of Grand Falls (Fig. 35c,d), sutures developed in the wasting ice sheet resulting in an abundance of interlobate morainal and glaciofluvial complexes. In a few places northward-oriented valleys were dammed by the retreating ice and short-lived lakes and ponds formed. Meltwater channels (Fig. 36) and morainal ridges suggest that glacier lobes, which developed as local topography began to influence the thinning ice sheet, may have interacted in a complex fashion, with ice from one lobe flowing over ice-free ground that had been previously overriden by an adjacent lobe. This phenomenon may explain some sequences north of Grand Falls where thin tills overlie washed sediments (Gauthier, 1983) and lacustrine sediments lie on hillslopes (Lee, 1959b).

When the ice margin stood at Grand Falls, meltwater probably graded to Inland Sea Acadia at about 130 m elevation (Fig. 35c). The lower stratified gravels in Saint John River valley below Grand Falls may relate to this time.

Western Sector. In the late part of the Plaster Rock phase the Kedgwick Highlands and northern Madawaska Hills were deglaciated except for areas covered by ice lobes in Green River valley above First Lake and in Iroquois River valley (Fig. 35c). The large lakes along the course of Green River are dammed by moraines formed by this glacier (Thibault, 1979). Interlobate ice-contact deposits were formed between the margin of North Maine ice retreating to the east and of Notre Dame ice retreating to the north adjacent to Iroquois River valley north of Edmundston.

The ice margin of the wasting North Maine Ice Divide may have paused near the southern edge of the Madawaska Hills just south of Quisibis Mountain as extensive ice-contact and morainal deposits occur along the slope here, most specifically the Quisibis Mountain moraine (39 of Fig. 20). In addition, meltwater channels are present that drained to the west and north (Fig. 35c). A large glaciofluvial complex was constructed in Quisibis River valley just north of Montagne-de-la-Croix by meltwater from one of these systems.

Shortly thereafter the North Maine Ice Divide disintegrated into a number of smaller ice centres; the St-Quentin Ice Centre became the main centre of ice in New Brunswick (Fig. 35c). Meltwater flowed subglacially between the Restigouche and Grand River systems north of Saint-Léonard during this stage, as is suggested by eskers along this divide. With the breakup of the North Maine Ice Divide, Saint John valley was deglaciated below Edmundston. Low ground was immediately submerged by Inland Sea Acadia into which the glacier was calving in Saint John River valley. Kiewiet de Jonge (1951) speculated that the DeGeer Sea (or Inland Sea Acadia here) may have been as high as 172 m near Grand Falls at its highest levels. As indicated in the section on lacustrine deposits, it certainly covered Grand Falls to an elevation of at least 158.5 m.

Quiet-water sediments in Saint John River valley above Grand Falls have always been attributed to glacial Lake Madawaska or Lake Madawaska, and this name has been retained for this arm of Inland Sea Acadia. This water body was certainly fed by glacial meltwater when ice was still present near Edmundston (Fig. 35d).

Remnant Ice Caps at Edmundston. Evidence suggests the presence of a late ice lobe in Madawaska and Saint John valleys near Edmundston (Fig. 35d). Lake sediments along Three Mile Brook near Ennemond (deposited in glacial Lake Ennemond; Fig. 22) indicate that late ice was present in Saint John River valley at the mouth of Three Mile Brook; glacial Lake Ennemond undoubtedly drained across a bedrock sill into Madawaska River valley. Faint southward-trending striae near Saint-Jacques and kame terraces inset into Madawaska River valley with southward-oriented paleocurrent directions also suggest that a lobe of ice, with a rapidly retreating southern margin (calving into Lake Madawaska), occupied this valley (Gauthier, 1982).

A lake may well have developed at this time to the north of the ice plug in Madawaska valley. High level deltas and beaches northwest of Saint-Jacques noted by Kiewiet de Jonge (1951) and Lougee (1954) may relate to this lake. Drainage of the lake would be either via the divide between Rivière Trois-Pistoles and Rivière Ashberish (Quebec) leading to St. Lawrence River and Lac Témiscouata respectively or the divide between Rivière Rimouski and Rivière des Aigles (Quebec) with similar drainage affinities (Fig. 33). The Rivière Rimouski outlet is the more probable one as it lies at a lower elevation (below 182 m elevation) than the Trois-Pistoles outlet. It is also marked by a fluvially cut channel, whereas the Trois-Pistoles/Ashberish divide is filled with ice-contact deposits. Alternatively, the high-level deltas and beaches northwest of Saint-Jacques could relate to a gradually rising water level caused by postglacial uplift of the Reversing Falls at Saint John. This alternative infers that the early phase of Lake Madawaska might have been controlled by the rock sill at Grand Falls rather than the Reversing Falls at Saint John.

St-Quentin and Sisson Branch Ice Centres. Meltwater channels on the central part of the Saint-Quentin Plateau generally indicate a shrinking ice body centred near the village of the same name. However, an east-west oriented, sharp-crested gravel ridge south of Sisson Branch Reservoir, similar ridges between Sisson Branch Reservoir and Tobique River, and the presence of lacustrine sediments around Sisson Branch Reservoir (Map 1594A) suggest that a small ice cap centred to the east of Sisson Branch Reservoir became separated from the St-Quentin Ice Centre in its later stages (Fig. 35d). The glacial lake, which covered the terrain around Sisson Branch Reservoir (glacial Lake Sisson, Fig. 22), was prevented from draining down a preglacial valley to the northeast of Sisson Branch Reservoir by this ice cap. Glacial Lake Sisson eventually drained via Sisson Brook after the course of present brook was established. Glacial Lake Sisson was fed by meltwater flowing from the St-Quentin Ice Centre through the low hills north of Sisson Branch Reservoir. Prior to the total decay of the Sisson Branch Ice Centre, the upper part of the Tobique River drainage was north via Oxford Brook (Fig. 35d).

When the southern margin of the St-Quentin Ice Centre retreated north of the hills to the north of Sisson Branch Reservoir, water was dammed between the ice and the hills. One such lake on Jardine Brook was partially filled with proglacial outwash. The final disintegration of the St-Quentin Ice Centre may have been quite complex with a number of small ice caps controlling the deposition of the glaciofluvial materials. For example, to the east and north of Saint-Quentin, crossbedding within glaciofluvial deposits suggests that the meltwater depositing these materials was flowing either to the east or west (present drainage is to the north).

Figure 37. *Rhythmically bedded clay under peat along Green River; alluvium overlies peat.*

Chronology

The Chaleur phase is well dated with marine shells underlying diamicton assigned to the Chaleur phase re-advance dated at 12 500 ± 130 BP (GSC-3552) and 12 300 ± 220 BP (GSC-3485; Table 3). Numerous shell dates are also available from either glaciomarine or marine sediments overlying Chaleur phase drift (Fig. 19, Table 3). These dates suggest that Chaleur ice was near its maximum position around 12.4 ka.

Towards the upper end of Baie des Chaleurs, dates are generally younger than 12.2 ka and ice may have persisted there until around 12.1 ka as suggested in Figure 20.

Dates on basal gyttja from lakes and bogs give a minimum age of 11.2 ka (Fig. 19) for deglaciation of terrain covered by Plaster Rock ice. Other older radiocarbon dates from postglacial sediments on terrain covered by Plaster Rock ice may be anomalously old: GSC-657 (13 800 ± 170 BP; Table 3) from southwest of Grand Falls is from marl associated with gyttja dated much younger; GSC-2557 (12 400 ± 110 BP; Table 3) from Lac Unique (Fig. 19) may be the result of the hard water effect or other contaminants (R.J. Mott, personal communication, 1982). Kite and Borns (1980) have counted 1200 rhythmites under peat, dated ca. 10 ka, at Green River (Table 3; Fig. 23, 37), but whether these are true varves is questionable. Thus the Plaster Rock phase is believed to date from between 12.4 and 12 ka. Perhaps the Lac Unique date of 12 400 ± 110 BP (GSC-2557) has some validity.

Madawaska Phase

The Madawaska phase refers to that time interval during which Saint John River valley above Grand Falls was free of ice, but was occupied by a major lake.

Lake Madawaska

Lake Madawaska has been traditionally considered to be the lake that formed extensive beaches and terraces in Saint John River valley between Grand Falls and Edmundston (Chalmers, 1885; Kiewiet de Jonge, 1951; Lee, 1959a; Kite and Borns, 1980) and was dammed by drift near Grand Falls (Chalmers, 1884; Kite and Borns, 1980) or glacier ice (Gauthier, 1980a; Gauthier and Thibault, 1980) during the Late Wisconsinan. As explained previously in the section describing the origin of Lake Madawaska, the lake probably owes its existence to differential tilting between Saint John and this area. For at least part of its existence, Lake Madawaska was only an arm of Inland Sea Acadia. Lake Madawaska was isolated to Saint John River valley above Grand Falls (Fig. 35d) when postglacial emergence caused the water level to fall below the level of the rock sill at Grand Falls (elevation slightly below 146 m). This initiated erosion of the Grand Falls gorge and development of the falls.

Lake Madawaska should have been deepest at its inception as isostatic depression was greatest at that time. If the rate of uplift at the supposed outlet at Saint John, however, was greater than the rate of uplift in the Grand Falls-Edmundston area following the lake's inception, some beaches would have formed later at higher levels than the earliest-formed beaches. This seems an unlikely scenario as postglacial uplift generally decreases exponentially with time following deglaciation.

Beaches would be best developed when lake levels were most stable, i.e., late in the history of Lake Madawaska when lake levels were controlled by slow erosion of the rock sill at Grand Falls. In the very early stages of Lake Madawaska, the rise and fall in water level may have been too rapid to establish beaches. Beaches are also not a common feature associated with the DeGeer Sea in either New Brunswick (Kiewiet de Jonge, 1951) or Maine (Leavitt and Perkins, 1935), probably because of continuously changing water levels. Kidson (1982) has also noted the difficulty of establishing former sea levels.

The Madawaska phase probably began ca. 12.2 to 12 ka at the end of the Plaster Rock phase. Later stages of Lake Madawaska predate 9.9 ka as indicated by the large number of radiocarbon dates obtained from peat overlying lacustrine clays at Green River, New Brunswick, and at Parent, Maine (Kite and Borns, 1981). Dates from Green River indicate that the lake had fallen to an elevation of 135 m by 9.9 ka. Rapid alluviation, evidenced by peat buried by alluvium and dated at 9285 ± 70 BP (SI-3706) at Green River and 8850 ± 65 BP (SI-3704; Table 3) at Parent, suggests that infilling of small lakes along Saint John River valley between Parent and Grand Falls had been completed by 8.8 ka.

Sea Levels

The Madawaska phase was generally marked by falling water levels in upper Saint John River valley. However, with the continued high relative level of the Reversing Falls, water elevations were probably above present ones along lower Saint John River. Hydrographic charts indicate a possible delta at 1.5 to 2 m below sea level at the mouth of Kennebecasis River and possibly at 3 to 4 m below sea level at the south end of Grand Lake. If drainage was to the south, the construction of these features requires the Reversing Falls to be lower than they are today and relative sea level to be below its present level at Saint John. Most probably the deltas formed at the time Inland Sea Acadia drained to the north through the Rimouski or Trois-Pistoles outlet (Fig. 33) north and west of Lac Témiscouata, Quebec.

Water levels were still relatively high between 11 and 10 ka in the middle reaches of Saint John River valley near Woodstock, as macerated wood and peat in clay at 105 m elevation near Woodstock were dated at 11 300 ± 200 BP (BGS-688) and 10 070 ± 150 BP (BGS-665; Table 2), respectively. Peat under clay on a ridge at 135 m elevation northwest of Woodstock was dated at 10 180 ± 150 BP (BGS-663).

No shell dates between 12 and 10 ka are available from southern or eastern New Brunswick and it can be assumed that sea level was below its present level throughout that time. Sea level at Saint John was probably 50 m lower than present between 12 and 10 ka (Fig. 24). Amos et al. (1980) showed that sea level throughout that time in the Bay of Fundy was below modern sea level. Kranck (1972) indicated that between 12 and 10 ka sea level was lower than present over much of Northumberland Strait and adjacent Gulf of St. Lawrence. Kranck's (1972, Fig. 6, p. 842) sea level curves, however, indicate that some low areas on the New Brunswick Lowlands may have remained below relative sea level at this time.

A shell date of 11 900 ± 100 BP (GSC-3450; Table 3) at Dickie Cove on Baie des Chaleurs (Fig. 19) at 13 m elevation indicates relative sea level there exceeded this elevation at that time. Continued postglacial uplift probably caused sea level to fall below the modern level in the bay shortly after as this is the youngest shell date from that area.

Late Wisconsinan and Early Holocene Climate

Late Wisconsinan ice centres in Maine and New Brunswick do not appear to have developed in mountainous or highland areas. Rather, they appear to have developed relatively independent of topography as noted by Kite et al. (1982); for example, the Escuminac and Gaspereau ice centres were located in lowlands. Kite et al. (1982) believed that the ice caps were "climatically dead", that winter storms tracked south of the ice caps, and that relatively dry climate existed over them. However, the build up of the Escuminac Ice Centre, Gaspereau Ice Centre, and Eastern Miramichi Ice Fields, and the vigorous flow of ice from these ice caps would suggest that they were nourished by winter storms tracking directly over their centres and that they were not "climatically dead".

Rapid deglaciation that occurred in the coastal areas of New Brunswick may have been caused by the effects of rising sea levels and chilling of ocean temperatures by icebergs and meltwater rather than general climatic warming (Ruddiman and McIntyre, 1981). The rising sea level would cause rapid calving and a general drawdown of the ice caps, whereas a cooler ocean would reduce the amount of snowfall received by the ice caps.

Prior to 12.6 ka, unglaciated parts of New Brunswick were covered by treeless tundra (Mott, 1975b). Along the coast a herbaceous tundra appears to have dominated, whereas inland some birch may have been present (Terasmae, 1973; Mott, 1975b). Modern analogs to pollen assemblages of this age are not available but the indicated cool climate is compatible with the co-existence of extensive glacier ice and ice-choked seas.

The climate began to warm around 12.6 ka with the continued retreat of ice. During this time, poplar and dwarf birch were succeeded by spruce along the southern coast (Mott, 1975b). This spruce forest suggests that a climate similar to that of present-day north-central Quebec existed in southern New Brunswick from about 12.6 to 11 ka.

A more open forest may have developed between 11 and about 9.5 ka (Mott, 1975b). Birch and other species increased in abundance at the expense of spruce (Picea). This may have been due to a short interval of climatic cooling.

Holocene

The Holocene was marked primarily by continued crustal uplift in central New Brunswick and by submergence in coastal areas where sea level rise exceeded uplift or where subsidence occurred as a peripheral glacial forebulge migrated inland (cf. Clark et al., 1978; Quinlan and Beaumont, 1981). This resulted in landscape changes along river valleys, mainly the Saint John, and along the coast.

River Regimes and Alluviation

Continued crustal uplift in northern and central New Brunswick resulted in the initiation of the falls on Saint John River at Grand Falls and the eventual filling of the bottom of Lake Madawaska with alluvium. Between Edmundston and Lac Témiscouata (Quebec), a sand plain was formed in upper Saint John valley (Kiewiet de Jonge (1951) and Lee (1955) have noted northward dipping crossbeds in the sandy terraces along Madawaska River). Kiewiet de Jonge (1951, p. 101, 102) suggested that this sand plain was formed as follows: "A sand laden Saint John River was pressing hard on the heels of vanishing Lake Madawaska and its deposits finally split the lake in two parts thus initiating the initial high stages of Lake Temiscouata. In flood stage the St. John River could pour part of its waters and sediments into the Madawaska Valley, raising the level of Lake Temiscouata and at the same time building a delta-like deposit as far north as Ste. Rose. By the time this condition was terminated by St. John River entrenching itself upon a southeast course (a condition resulting from land tilting or from down cutting at Grand Falls, or both), the completed sand plain had filled the Madawaska Valley over a distance of 21 miles, flush from side to side, and its deltaic terminus at Ste. Rose determines the southern limits of Lake Temiscouata today." Between Edmundston and Grand Falls the lake bottom was filled in by a meandering river flowing over a relatively flat surface. Although alluviation may have begun around 10 ka in this part of Saint John River valley, it was most active between 9 and 4 ka (Kite et al., 1982). Following this, some stream incision occurred, although the major Siegas Terrace was still subject to flooding at 815 BP (Kite et al., 1982).

As the water level regressed below Grand Falls, terraces were developed in the following manner (Rampton and Paradis, 1981a, p. 33): "It is thought that as emergence proceeded, the Saint John River continually eroded drift from its valley and deposited it at the head of an ever-lowering and receding estuary. As the head of the estuary moved downstream the river would continue to entrench itself, erode previously deposited alluvium and form lower terraces. The general stratigraphic sequence of sand and gravel over clay in the valley bottom and presence of terraces along the whole valley would seem to corroborate a model of alluvium being deposited in a migrating delta front over estuarine sediments being subsequently eroded to form the terraces".

In Saint John River valley below the mouth of Keswick River, the series of terraces outlined above is absent. Instead, the valley sides within the limits of Inland Sea Acadia are veneered by sand, resulting from wave and current action (Seaman and Thibault, 1981). Only a low-level terrace, comprising sandy alluvium overlying deepwater silts and clays, is present. Freshly exposed sandy areas were subjected to eolian activity, with subsequent incision of the river into this terrace, and sand dunes developed, especially near the mouth of Oromocto River.

Coastal submergence has resulted in drowning of lower Saint John River valley and the extension of tidal effects to the mouth of Keswick River. Here a "delta-like distributary system of channels and islands" (Seaman, 1982b, p. 88) has resulted. Extensive flat alluvial plains with numerous lakes, distributary channels, and levees along lower Saint John River valley and tributary valleys have developed because of this resubmergence.

Other river valleys have been affected in similar fashion to the Saint John. Isostatic uplift has resulted in erosion by streams in their headwaters and deposition of alluvium over lacustrine and marine deposits along their lower courses as water levels regressed. Coarse alluvium overlying clay and silt is common along the lower course of most major streams. The lower parts of these floodplains and terraces have generally been flooded during late Holocene submergence. At present, streams are generally building small deltas at the coast in an environment of quasi-equilibrium.

Coastal Events

Maximum Emergence. Maximum emergence of the land and adjacent seafloor along the eastern coast occurred about 7 ka (Kranck, 1972; Prest, 1973). Amos et al. (1980) indicated that maximum emergence of about 25 m occurred in the Bay of Fundy between about 10 and 8 ka, whereas Schnitker (1974) indicated maximal emergence of 65 m ca. 8.5 ka.

Coastal Submergence. Numerous dates on stumps and terrestrial peat below the present limit of high tide (Tables 2, 3) indicate that relative sea level has risen over the last 4 ka along the Bay of Fundy and Northumberland Strait (cf. Grant, 1970; Prest, 1973; Rampton and Paradis, 1981b, c). For example, rooted stumps 10 to 12 m below high tide near Fort Beausejour east of Sackville dated at 4040 ± 130 BP (GSC-930) and 4120 ± 130 BP (GSC-1089; Table 2).

Grant (1970) has indicated that relative sea level has risen recently at a rate of 0.15 m per 100 years in the Gulf of St. Lawrence, and 0.30 m per 100 years in the Bay of Fundy. D. Keys (personal communication, 1983) has concluded from an analysis of radiocarbon dates on peat from near sea level on the Acadian Peninsula (BGS-518, -519, -520, -521, Table 3), from the elevation of the dated material relative to sea level, and from the rate of peat accumulation that sea level is presently rising between 0.11 and 0.18 m per 100 years along the eastern Acadian Peninsula.

Present coastal subsidence is slow enough, albeit continuous, that salt marshes persist in sheltered coastal environments. Intertidal sediments continue to accumulate in those environments and on large intertidal mud flats. The latter are most extensive along Chignecto Bay where tidal amplitudes are greatest.

Slowly rising sea level has contributed to coastal retreat and to the development of large spits and baymouth bars. Relatively friable sandstone along Northumberland Strait and Gulf of St. Lawrence is allowing coastal recession to proceed at a rate of 0.3 to 0.7 m per year (Airphoto Analysis Associates Consultants Limited, 1975a). This has resulted in large sand dunes reported by Ganong (1906a) at the southwest corner of Miscou Island being completely eroded. Coastal recession along the Bay of Fundy is generally less than 0.25 m per year (Hunter and Associates, 1982) because of the resistant bedrock present; only in areas where unconsolidated sediments occur does coastal recession exceed 0.75 m per year. Locally coastal erosion has led to slope failures along the coast (Ruitenberg et al., 1976; Ruitenberg and McCutcheon, 1978; Hunter and Associates, 1982).

Large spits and bars are best developed where the tidal amplitude is lowest and where the seafloor slopes gently away from the land. These characteristics, which are typical

of New Brunswick's eastern shoreline, have allowed large wedges of sand to be moved shoreward with time as coastal submergence and recession occurred. Wind has continued to affect the large sandy spits and baymouth bars, creating sand dunes along their inland edges; most spectacular are the stabilized and presently active dunes at the north end of Miscou Island.

Terrestrial Events

Weathering and Soil Development. Since the landscape has been uncovered by Late Wisconsinan glaciers and Holocene water bodies, its surface has been subjected to weathering and soil formation (Rampton and Paradis, 1981a, b, c). On the Northern Miramichi Highlands exfoliation and weathering of the granites have occurred during the Holocene. An unvegetated skirt of grus is present immediately adjacent to most granitic tors suggesting relatively recent activity, although Gauthier (1980b) has noted a siliceous glaze on some grus, which he believed indicates some antiquity of the grus deposition. Well developed podzols are present on the grus on some slopes and attest to its recent stability. Weathering pits (Fig. 38), some showing recent activity, are present on the flat upper surface of some tors.

Karst topography has developed in a few areas in the Aganance Ridges, Shepody Ridges, and Tobique Basin where gypsum and anhydrite occur (Rampton and Paradis, 1981a, c). The karst likely developed mainly during the Holocene, but a sinkhole near Hillsborough contains peat dated at > 43 000 BP (GSC-1680) and a mastodon bone dated at 13 600 ± 220 (GSC-1222; Table 2) in a sequence of silts (Schroeder and Arsenault, 1978).

Holocene Climate. Climate was warming in New Brunswick between 10 and 8.5 ka as indicated by increasing **Pinus** pollen in pollen sequences (Mott, 1975b; Korpijaakko, 1976). Pollen sequences indicate a continued warm climate between 8.5 and 3 ka with fluctuations in the moistness of the climate. This period of warmth is coincident with the general worldwide warm conditions of the Hypsithermal, but is partly due to continued low relative sea level along the coast of New Brunswick. The shoreline, which was offshore from its present position, resulted in restricted circulation in the Bay of Fundy and in a more continental climate in New Brunswick.

Climate began to cool some time between 4.5 ka (Korpijaakko, 1976) and 3 ka (Mott, 1975b). During the last 2 ka, climate appears to be also marked by an increase in humidity. These climatic changes may be due either to general climatic fluctuations or to rising relative sea level along the coast of New Brunswick.

Peatland Development. Organic deposits began to accumulate (cf. Table 2, 3), mainly in depressions, following the exposure of land from under sea and ice in New Brunswick. Even in flat areas, however, poor drainage and impervious soils under a relatively humid climate eventually led to paludification of the landscape and the development of peatland. Peatland development was initiated or intensified between 8.5 and 8 ka over most of the province (Korpijaakko, 1976).

Dated peat profiles (Tables 2, 3) indicate that the rate of peat accumulation has increased recently both in the southern (Rampton and Paradis, 1981c) and eastern (D. Keys, personal communication, 1983) parts of the province. This could be either a function of a more maritime climate or a function of bog succession.

Tectonic Events

Tectonism is New Brunswick through the Holocene has been confined mainly to differential crustal subsidence and minor faulting and earthquakes.

Matthew (1894c) noted a number of displacements, which he attributed to faulting, on glaciated surfaces at Saint John. Whether this is due to tectonic thrusting from the southeast, as Matthew envisaged, or crustal relaxation following deglaciation is not known. Minor displacement of striae on slate has been observed near Edmundston. The only other reported fault displacement of glacial deposits is from along the Oak Bay Fault on Campobello Island (Rast et al., 1979).

Grant (cf. 1980a) has not only documented a rising sea level around the coast of New Brunswick, but has shown that differential crustal subsidence is occurring in New Brunswick at present. For example, the Sackville area has subsided about 12 cm more than the Fredericton area and 17 cm more than the Edmundston area over the 50 years between 1910 and 1970. W.B. Thompson (personal communication, 1983) also believes that the Passamaquoddy Bay region is presently subsiding more rapidly than surrounding areas.

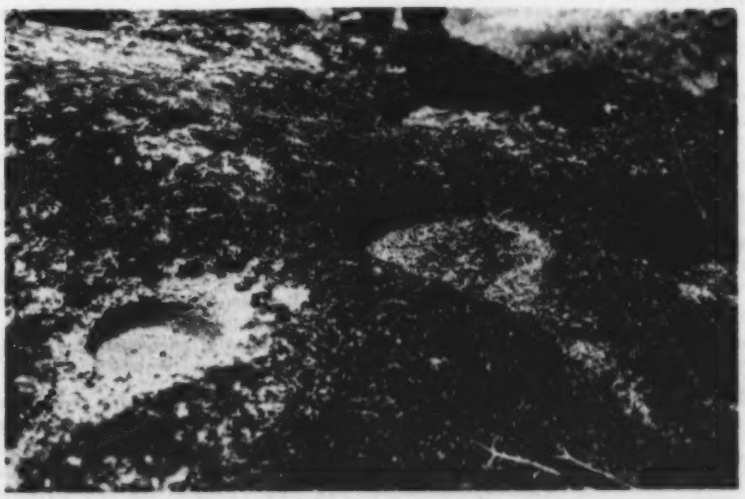


Figure 38. *Weathering pits in a tor.*

Earthquakes, generally minor, are common in New Brunswick. They appear to have a random distribution except for some concentration along the Oak Bay Fault in southwestern New Brunswick (Rast et al., 1979). Generally, historically reported earthquakes show a reduction in intensity. The most recent major earthquake occurred in 1982 near Trousers Lake in the Central Miramichi Highlands.

CONCLUSIONS

1. The basic topography of New Brunswick results from pediplanation (or peneplanation) during the Cretaceous and Tertiary and the subsequent uplift and dissection of these plains during the late Tertiary and Quaternary.
2. Weathered bedrock in the subsurface throughout the province and at the surface on the Northern Miramichi Highlands was mostly weathered prior to Late Wisconsinan time and, in many places, prior to Wisconsinan time.

3. Tors, periglacial features, and development of mature soils on the Northern Miramichi Highlands indicate that this area was glaciated only prior to Wisconsinan time. Whether the highlands were ice-free during the Wisconsinan or whether they were covered by cold-based, noneroding glacier ice is a matter of controversy.

4. Intricately dissected terrain around the flanks of the Central Plateau of the Caledonian Highlands, weathered unconsolidated materials and periglacial features on its surface, and a striation pattern similar to that present on a nearby Sangamonian marine rock bench indicate that the plateau was glaciated during Early or Middle Wisconsinan time.

5. Weathered horizons on subsurface till and glaciofluvial deposits, organic materials beyond the range of radiocarbon dating, oxidized striated surfaces, boulder trains, and striation patterns unrelated to Late Wisconsinan striation patterns were all developed during glacial and nonglacial intervals prior to the Late Wisconsinan.

6. During the Chignecto phase of the Late Wisconsinan, estimated to have occurred between 18 and 15 ka ago, ice flowed down Bay of Fundy and onto the Acadian Peninsula from the Escuminac Ice Centre in the Gulf of St. Lawrence. Major glaciomarine deltas formed adjacent to the Bay of Fundy as the glacier ice retreated from this area.

7. Beaches and hillside patches of clay and silt over 70 m above sea level and well above the level of deltas formed during retreat of Chignecto ice define a high stand of sea level along the Bay of Fundy that occurred near the end of the Chignecto phase.

8. Glacial flow patterns in eastern New Brunswick indicate that a large ice cap, the Gaspereau Ice Centre, strongly influenced ice flow in eastern New Brunswick between ca. 13.4 and 12.7 ka during the Bantalor phase. Glacial flow along the eastern and northern edges of the Gaspereau Ice Centre was affected by ice flowing from the North Maine Ice Divide and from ice centres in the Notre Dame Mountains.

9. Fossiliferous till near Saint John dated ca. 13.2 ka indicates an active glacier ice margin terminating in the sea at that time near Saint John.

10. The elevation of delta surfaces along the Bay of Fundy west of Saint John indicates that relative sea level rose shortly after 13.2 ka and then fell continuously throughout the remainder of the Late Wisconsinan and early Holocene.

11. Easterly trending esker segments and striae along former ice margins in southeastern New Brunswick indicate that Gaspereau ice retreated slightly more quickly than ice from the North Maine Ice Divide during deglaciation of southwestern New Brunswick.

12. Probable submergence of much of the Central Lowland and adjacent areas to relatively high levels suggests that these areas were deglaciated before 13 ka when water levels were probably at their maximum elevation in these areas.

13. Areas showing no definite sign of submergence in the New Brunswick Lowlands probably mark the site of remnant ice caps that persisted to near the end of the Bantalor phase.

14. Residual isostatic depression of the New Brunswick Lowlands, and relatively low sea levels in and the bathymetric configuration of the Gulf of St. Lawrence may have led to a large fresh or brackish water body covering the New Brunswick Lowlands during part of its deglaciation.

15. Varved sediments near the mouth of Sevogle River attest to the presence of a long-lived glacial lake – Sevogle – in their vicinity.

16. A major belt of ribbed moraine along the western edge of the Southern Miramichi Highlands and St. Croix Highlands defines the maximum extent of ice flowing south along terrain adjacent to Saint John River valley in west-central New Brunswick during the Millville/Dungarvon phase ca. 12.7 ka. The margin of ice in northern New Brunswick during this time is poorly defined.

17. Some of the terrain covered by ice during the Bantalor phase and Millville/Dungarvon phase in western New Brunswick is characterized by large esker and valley train glaciofluvial systems and by belts of ribbed and ridged moraine. The latter indicates that the glaciers were at least sporadically active near their margins.

18. Numerous short-lived glacial lakes formed in valleys blocked by retreating ice margins, e.g. glacial Lake South Oromocto, and possibly by crustal warping, e.g. glacial Lake Napadogan, during the Bantalor and Millville/Dungarvon phases.

19. Saint John River valley was flooded by DeGeer Sea after retreat of Bantalor and Millville/Dungarvon ice. Major deltas were deposited within the valley near the retreating ice margin.

20. Shells dated ca. 12.4 ka from the base of a diamicton along Baie des Chaleurs indicate an easterly re-advance of glacier ice along the bay at this time. A major interlobate moraine was deposited near sea level upon decay of this lobe.

21. Between 12.4 and 12 ka, during the Plaster Rock phase, the North Maine Ice Divide and Notre Dame Mountains Ice Centres broke up into a number of independent small ice caps and wasted away. This phase was marked by the formation of interlobate glaciofluvial deposits and ablation deposits.

22. Drainage along the eastern edge of the ice cap in northern New Brunswick during the latter part of the Millville/Dungarvon phase and the Plaster Rock phase was diverted northward into the Upsalquitch River drainage system where a glacial lake – Upsalquitch – formed because of ice in Baie des Chaleurs.

23. Two small ice caps on the Saint-Quentin Plateau, the St-Quentin and Sisson Branch ice centres; a small ice mass near the junction of Saint John and Madawaska rivers; and small ice caps in the Notre Dame Mountains mark the last remnants of glacier ice in northwestern New Brunswick.

24. Upper Saint John River valley was flooded by the DeGeer Sea initially upon deglaciation during the Plaster Rock phase. Later this water body, Lake Madawaska, became part of Inland Sea Acadia when the Reversing Falls at Saint John rose above sea level ca. 12.3 ka.

25. Lake Madawaska became an entity separate from the remainder of Inland Sea Acadia when isostatic uplift exposed the bedrock sill at Grand Falls upon which the falls developed. Lake Madawaska probably drained to St. Lawrence River via Lac Témiscouata and Rivière Rimouski for part of this interval.

26. Radiocarbon-dated peats overlying Lake Madawaska rhythmically bedded clays indicate that much of Lake Madawaska was infilled with sediment by about 10 ka. Inland Sea Acadia persisted below Grand Falls for some time following this as minimum relative sea level at Saint John was not achieved until ca. 8 ka.

27. Isostatic uplift during the Holocene led to the alluvial capping and the terracing of lacustrine and marine deposits along the lower part of many river valleys.

28. Peatlands began developing throughout New Brunswick following deglaciation and emergence of the land surface, but peatland formation intensified after 8.5 ka. The rate of peat accumulation has increased during the late Holocene.

29. Dated stumps and terrestrial peat underlying intertidal sediments indicate that submergence of the New Brunswick coastal areas has been occurring continuously since at least 4 ka.

30. Differential crustal subsidence is presently occurring in New Brunswick.

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APPENDIX I

Description of Physiographic Divisions and Subdivisions
of New Brunswick (cf. Fig. 1)

A1 – EDMUNDSTON HIGHLANDS

Physiography

The Edmundston Highlands consist of hilly terrain with summits rising from 360 m elevation near Madawaska River to over 580 m near their northern edge. The highlands exhibit more relief and a more rugged aspect than the Chaleur Uplands to the east from which they are separated by a discontinuous, 60 to 90 m-high escarpment. The Edmundston Highlands have been subdivided into the Saint-François Hills, Madawaska Hills, and Kedgwick Highlands.

The Saint-François Hills lie south of Madawaska River and consist of ridges and valleys, many with a northwest-southeast orientation, probably produced by glacial fluting. Ridges and hills are moderately steep-sided with broad crests. Hill crests are higher at the southeastern edge (maximum elevation 532 m) than near Madawaska River (where they fall to below 425 m). Relief within the hills is commonly between 150 and 210 m, and the area is well drained except for a few broad valleys. A subparallel drainage pattern has developed in which the lower courses of most streams reflect the northwest-southeast orientation of the ridges and valleys. All streams drain to Saint John River.

The Madawaska Hills, which lie north of Madawaska River, consist of hills and ridges whose summits generally do not exceed 460 m elevation. Summit areas are generally broad. Valleys are relatively narrow except for the valleys of Iroquois and Green rivers which drain the area. Hill crests range between 360 and 390 m elevation along Madawaska River valley, and are between 430 and 460 m elevation in the central and northern areas. Relief is commonly 120 to 150 m in the south and 150 to 215 m in the north. The steepest slopes and greatest relief occur along the deeply incised major rivers. The area is well drained except for some parts of wider valleys. Most streams have a broadly dendritic drainage pattern, although locally segments have a north-northeast to northeast parallel pattern. The latter is probably a reflection of glacial fluting.

Madawaska River valley is a broad (0.7 to 1.0 km wide), flat-bottomed valley which transects the northeast-southwest axis of the Edmundston Highlands.

The Kedgwick Highlands are the highest part of the Edmundston Highlands with most hill and ridge crests above 455 m elevation; the highest peak is at 604 m. The western and northern parts consist of rolling hilly terrain with a few broad depressions and valleys which are poorly drained where they form the headwaters of streams. Relief generally varies between 90 and 120 m; it approaches 185 m along the deeply incised Green River valley. The southeastern part has greater relief due to stream incision. Relief of 210 to 245 m is common along most minor streams and between 300 and 365 m along the deeply entrenched Patapedia and Kedgwick rivers. The area is well drained with broad ridges separated by narrow, commonly V-shaped valleys. Drainage patterns in the Kedgwick Highlands are complex; changing from a partially deranged dendritic pattern in the western part to a northwest-southeast oriented parallel drainage pattern in the deeply dissected eastern part.

Bedrock Geology Summary

The Edmundston Highlands lie totally within the Edmundston zone (Fig. 2) of the Gaspé Synclinorium and they are underlain by grey micaceous slate and greywacke of the Temiscouata Formation (Ruitenberg et al., 1977). Small granite intrusions are present along the southeastern boundary (Potter et al., 1979).

A2 – CHALEUR UPLANDS

Physiography

The Chaleur Uplands consist primarily of a plateau, locally dissected near major streams and interrupted by clusters of hills. The uplands, although generally forming a plateau well above sea level, are lower and less rugged than adjacent ground in the Edmundston Highlands and New Brunswick Highlands (Rampton and Paradis, 1981a; Gauthier, 1983). The narrow, but low, coastal plain adjacent to Chaleur Bay has been included as part of the Chaleur Uplands because of its geographic position and its location within the Tobique-Chaleur Synclinorium, which underlies the northeastern part of the Chaleur Uplands (Fig. 2).

The Williamstown Plateau flanks Saint John River in the west-central part of New Brunswick. Rampton and Paradis (1981a, p. 4) previously described this area as "a gently rolling upland with relief between 30 and 60 m (100 and 200 ft) except where the Saint John River Valley and its tributaries are incised up to 150 m (500 ft) into it"; a few isolated hills exceed 250 m elevation. The area is well drained except for valleys and depressions whose outlets are clogged with drift. Tributaries to the Saint John have dendritic drainage patterns showing only minor effects of glacial derangement (Rampton and Paradis, 1981a).

The southwestern boundary of the Kintore Hills is marked by a bedrock-controlled scarp between 150 and 245 m in height. The Kintore Hills are underlain primarily by sandstone and volcanic rocks whereas the adjacent Williamstown Plateau and Drummond Uplands are underlain by calcareous slates and limestones (St. Peter, 1978; Fyffe, 1982b). Local relief generally varies between 120 and 245 m, but is locally up to 270 m along Saint John River and in excess of 300 m in the Cameron Mountain Range southwest of Grand Falls (one summit is at 572 m elevation). The Kintore Hills merge imperceptibly with the Miramichi Highlands to the east. The area is well drained due to the relief and steep slopes. A dendritic drainage pattern has developed.

Tobique Basin, primarily underlain by Carboniferous sediments, has a flat to gently sloping surface with little relief except near its western edge, where it is gently rolling, and where bedrock-controlled mesa and cuesta scarps between 60 and 150 m high are present. The western part of the basin is moderately well drained due to the relief, but drainage is fair to poor in the eastern part. The drainage pattern is dendritic to slightly irregularly dendritic.

The Drummond Uplands flank Saint John River between the mouths of Aroostook and Grand rivers. The surface of the Drummond Uplands is rolling with local relief of 30 to 60 m. Most hill crests are below 270 m elevation. Tributaries of the Saint John are incised up to 120 m into the upland surface, and the landscape has the appearance of a dissected plateau. The drainage pattern is dendritic and the area is well drained due to its rolling nature and stream incision.

The Siegas Hills lie north of Saint John River between Grand and Green rivers. This area consists of steep-sided hills and ridges separated by broad, commonly poorly drained valleys. The ridges and hills result from dissection of the

plateau surface (present to the north) by the lower courses of streams tributary to Saint John River. The ridges and valleys have a northeast-southwest alignment, which reflects the underlying major structure (St. Peter, 1978).

The Central Saint John River Valley consists of the floodplain and low terraces and benches that border Saint John River between Edmundston and Grand Falls. The valley is broader in this segment – its width varying between 1.5 and 2.5 km – than up or downstream. Drainage is variable according to topographic position relative to the river.

The Saint-Quentin Plateau forms the core of the Chaleur Uplands in northwestern New Brunswick. For the most part it has an undulating surface with broad depressions centred near Saint-Quentin and Sisson Branch Reservoir and a broad trough paralleling segments of Grand and Restigouche rivers. These low areas are generally underlain by Matapedia Group limy rocks. West of Restigouche River the ground slopes gently up to the Notre Dame Mountains. Most of the plateau lies between 270 and 370 m elevation, has an undulating surface, and has local relief between 30 and 60 m. West of Sisson Branch Reservoir a group of hills are present with peaks to 520 m elevation and relief of 90 m. A few steep-sided hills (Blue Mountain and associated peaks) rise 210 m to elevations of about 545 m in the southeast part. Most of the plateau is moderately well drained, except for a few poorly drained, low, flat areas. Many low ridges on the upland are parallel because of structural control of erosion

Figure 30. *Dissection and drainage pattern in siltstones adjacent to Tobique River near Riley Brook (NAPL 23859-90, 91).*

(Gauthier, 1983) and because of glacial fluting. Segments of Restigouche River and the Tobique and Mamozekel rivers flow in deep trenches paralleling structural trends and are incised 90 to 120 m below the upland surface. Some areas underlain by siltstones adjacent to Tobique and Mamozekel rivers are intricately and deeply dissected (Fig. 39). Drainage patterns of the above rivers are subparallel to rectangular, reflecting structural control. Upsalquitch River, lower course of Restigouche River, and lower courses of their tributaries are incised up to 150 m below the upland surface; these streams have developed a broadly dendritic drainage pattern with a rectangular aspect.

The Jacquet Plateau lies immediately northeast of the Saint-Quentin Plateau, Upsalquitch River being the arbitrary division. The plateau surface is undulating with relief generally less than 30 m and a few hills projecting 60 to 90 m. Its surface is well drained except for a few poorly drained broad depressions. Its southern edge is hilly with local relief of 75 to 100 m where it merges with the Miramichi Highlands. The plateau surface slopes to the north and east from elevations between 360 and 400 m along its southwestern and central parts. About 15 km from its northern edge, its slope increases slightly at 275 m elevation down to the 60 m level at its northern edge. Similarly, about 12 km from its eastern edge, its slope increases at 150 m elevation down to the 60 m level at its eastern edge. Drainage patterns are generally parallel as glacial flow and erosion paralleled the east-northeast and northeast structural trends in this area. Tributaries of Upsalquitch River and the middle and lower segments of Charlo, Jacquet, and Tetagouche rivers are incised in V-shaped valleys, 75 to 180 m below the upland surface.

The Campbellton Hills are a group of structurally controlled, steep-sided ridges and hills lying along the northern edge of the Saint-Quentin Plateau. They are underlain primarily by basalts, andesites, and calcareous rocks of the Devonian Dalhousie Group, whereas the immediately adjacent parts of the Jacquet Plateau are underlain by calcareous rocks of the Silurian Chaleur Group. Much of the terrain lies above 300 m elevation. The elevation of ridge crests declines in an eastward direction with a maximum of 483 m near the western edge and just over 180 m near the eastern edge. Local relief is generally 150 m in the west and between 60 and 90 m in the east. The area is well drained with streams generally defining a subparallel to parallel drainage pattern reflecting structural control. Some glacial diversion of drainage has occurred.

The Chaleur Coastal Plain is a gently sloping to slightly undulating plain that borders Baie des Chaleurs. Most of the coastal plain lies below 70 m elevation where a marked steepening in slope occurs as the terrain rises to the core of the Chaleur Uplands. All streams rising in the adjacent Jacquet Plateau and Saint-Quentin Plateau transect the coastal plain. The plain is moderately well drained. Coastal cliffs are generally less than 10 m high.

Bedrock Geology Summary

The Chaleur Uplands are primarily underlain by rocks comprising the Matapedia-Aroostook Anticlinorium, Tobique-Chaleur Synclinorium, and Elmtree Dome (Fig. 2). Each tectonostratigraphic zone is characterized by a different suite of rock types.

The Matapedia-Aroostook Anticlinorium, which underlies the western part of the Chaleur Uplands, is characterized by calcareous shales and slates and argillaceous limestones of the Matapedia Group of Ordovician-Silurian age; and argillites, argillaceous sandstones and limestones, and greywackes of the Grog Brook Group of Ordovician age. The Silurian Perham Group, consisting mainly of quartzose sandstones, slates, siltstones, and limestones, occurs at its southern extremity (Fyffe, 1982a). These rocks were formed during the Taconian Orogeny and were subsequently folded and metamorphosed during the Acadian Orogeny (Poole and Rodgers, 1972; Ruitenberg et al., 1977).

The Tobique-Chaleur Synclinorium, which underlies most of the eastern part of the Chaleur Uplands, is characterized by argillaceous and calcareous siltstones and sandstones, rhyolites, andesites, and basalts of the Chaleur Group of Silurian age; and calcareous sandstones and siltstones, limestones, andesites, basalts, and tuffs of the Dalhousie Group of Devonian age (Ruitenberg et al., 1977). Syntectonic deposition, folding, and faulting occurred entirely during the Acadian Orogeny (Poole and Rodgers, 1972). The Elmtree Dome, present near the northeastern extremity of the Chaleur Uplands, is characterized by gabbros, basalts, quartzose greywackes, and slates of the Ordovician Elmtree and Fournier groups (Ruitenberg et al., 1977). These rocks have been intruded by Devonian granite (Davies, 1977). The Elmtree Dome contains a suite of rocks similar to that within the Miramichi Anticlinorium.

Carboniferous redbeds overlap the above sequences in the Plaster Rock basin west of Grand Falls (Fig. 2) and along Baie des Chaleurs. These rocks are primarily siltstones, sandstones, and conglomerates with minor limestone and gypsum (Davies, 1977; Fyffe, 1982a).

Major fault systems trend mainly northeast in the Matapedia-Aroostook Anticlinorium and southwestern part of the Tobique-Chaleur Synclinorium, but trend east-northeast in the northern parts of these tectonostratigraphic divisions. Fault-line scarps, linear troughs, and parallel stream courses are common geomorphic expressions of these faults.

A3 – MIRAMICHI HIGHLANDS

Physiography

The Miramichi Highlands comprise that portion of the New Brunswick Highlands lying north of Saint John River (Williams et al., 1972). The Miramichi Highlands are higher, more rugged, and hilly than the adjacent Chaleur Uplands and New Brunswick Lowlands. The West Bathurst Basin has been included as part of the Miramichi Highlands because of its location within the Miramichi Anticlinorium, which underlies much of the highlands (Fig. 2).

The Northern Miramichi Highlands comprise that hilly to mountainous part of the Miramichi Highlands lying north of Little Southwest Miramichi River. Gauthier (1983) has described this area as follows: "In this region are found the highest summits of New Brunswick. They form a central undulating high plateau with an average elevation well above 600 metres. Mount Carleton is the highest peak with an elevation of 820 metres."

This description applies mainly to the central core of the Northern Miramichi Highlands roughly outlined by Mount Carleton and Wilkinson Mountain on the west and Popple Depot and Big Bald Mountain on the east, where numerous rounded peaks are joined by broad ridges. Streams are deeply incised and the area is well drained with relief commonly between 210 and 250 m. Maximum relief at Mount Carleton is 490 m. East and south of this massif elevations generally decrease to between 420 and 550 m, and relief to between 90 and 150 m. In the southwest part of the Northern Miramichi Highlands, however, Gulquac, Black, and Serpentine mountains form ridges, which peak at elevations of 622, 681, and 613 m, respectively, and stand up to 200 m above the surrounding basins and hills. In contrast to the massive ridges and hills separated by narrow deeply incised

valleys of the north-central massif, the southern and eastern areas consist of ridges and hills separated by broad depressions commonly containing lakes and poorly drained ground. In the southeastern part of the Northern Miramichi Highlands ridges and valleys have a northwest-southeast orientation, probably a result of glacial erosion.

Nepisiguit River-Upsalquitch Lake valley crosses the Northern Miramichi Highlands. This superimposed drainage system is entrenched up to 370 m along most of its length. The present drainage courses may in part be the result of glacial diversions. The Nepisiguit River system originates at Nepisiguit Lakes, which are in a broad basin underlain by Devonian rock (Davies, 1977). Drainage patterns are generally dendritic except in the southeastern part where a subparallel pattern is present, and in the southwestern part where glacial activity has blocked northwest flowing streams and deranged the drainage.

The Central Miramichi Highlands lie between Little Southwest Miramichi and Southwest Miramichi rivers. These highlands are lower than those to the north, although a good part of them stand above 300 m and are higher than the Southern Miramichi Highlands. Rampton and Paradis (1981a, p. 4) described them as follows: "Adjacent to the Southwest Miramichi River, the Highlands consist of high broad ridges with elevations ranging from 365 m (1200 ft) to 455 m (1500 ft) and with intervening valleys and basins 90 to 150 m (300 to 500 ft) below the ridge crests; hills along the ridge crests are 30 to 90 m (100 to 300 ft) higher than the general elevation of the ridges. West and southwest of Tuadook Lake, hills of 550 to 670 m (1800 to 2200 ft) stand 180 to 245 m (600 to 800 ft) above the intervening valleys and basins and give the landscape a more hilly aspect ... Drainage throughout the Miramichi Highlands is good because of the relief, although valleys are locally poorly drained because of preglacial drainage patterns being deranged and in places blocked by drift." The drainage pattern is irregularly dendritic, typical of glacially deranged drainage on a dissected upland. Some streams have had their courses diverted eastward because of glacial retreat to the west. Narrow, steep bedrock walls along parts of Southwest Miramichi River and some of its tributaries may be due to the late Pleistocene origin of the river course here.

The Southern Miramichi Highlands lie between Southwest Miramichi River and Saint John River. In the southern part only isolated hills exceed 300 m elevation and the topography is characterized by rolling hills, broad depressions, and relief of 60 to 90 m. Relief is up to 180 m along bedrock-controlled scarps. The northern part is much hillier with summits commonly exceeding 300 m elevation and relief being between 90 and 150 m. The area is generally well drained, although the broader depressions and upland swales may only exhibit poor to fair drainage. Drainage patterns are similar to those of the Central Miramichi Highlands.

The Eastern Miramichi Highlands flank the eastern edge of the Miramichi Highlands and are a continuation of the gently rising plain that constitutes the adjacent New Brunswick Lowlands. Essentially, these highlands form a transitional zone between the hill and mountain core of the Miramichi Highlands and the plain to the east. South of Renous River the western boundary of the Eastern Miramichi Highlands parallels a fault-line scarp. The Eastern Miramichi Highlands generally lie below 365 m elevation, have a gently rolling surface on which relief rarely exceeds 60 m, and are moderately well drained. Major streams are commonly incised 90 to 120 m below the upland surface. "Streams along the eastern edge of the Miramichi Highlands have developed a drainage pattern paralleling the regional slope and glacial fluting of the area" (Rampton and Paradis, 1981a, p. 4).

The West Bathurst Basin, which is located at the northeast tip of the Miramichi Highlands and west of Bathurst Harbour, consists of flat to gently undulating terrain lying below 60 m. Geomorphically, it is more closely related to the adjacent Chaleur Coastal Plain and Curventon-Bathurst Valley than to the Miramichi Highlands. Unconsolidated sediments in the basin are generally much thicker than in adjacent physiographic subdivisions. It is generally well drained except in parts of the southeast.

Bedrock Geology Summary

The core of the Miramichi Highlands are underlain by rocks of the Miramichi Anticlinorium (Fig. 2), mainly the metamorphosed Tetagouche Group of Cambrian and Ordovician ages (Ruitenberg et al., 1977). The Tetagouche Group consists of paragneiss, schist, amphibolite, quartzite, quartz wacke, phylite, and slate overlain by volcanic rocks interbedded with slate, greywacke, iron formation, and limestone. These rocks were intensely deformed and metamorphosed during the Taconian Orogeny; some deformation occurred during the Acadian Orogeny (Ruitenberg et al., 1977). Extensive granitic intrusions are present in the sequence that dates from Ordovician to Devonian, possibly Mississippian (Fyffe, 1982a).

The western part of the Miramichi Highlands includes rocks of the Tobique-Chaleur Synclinorium (Fig. 2), mainly volcanic rocks of the Devonian Tobique Group (Davies, 1977; Fyffe, 1982a). Siltstones and sandstones are also present within this group. Felsic volcanics form most of the higher mountains, e.g. Mount Carleton along the western edge of the highlands.

The southeastern edge of the Miramichi Highlands is underlain by Silurian greywacke and slate located in the Magaguadavic Zone (Fig. 2; Fyffe, 1982a) or Fredericton Trough (Poole and Rodgers, 1972). These rocks have been cut by Devonian granitic plutons. The rocks were folded during the Acadian Orogeny.

A4 – ST. CROIX HIGHLANDS

Physiography

The St. Croix Highlands form a broad V around the apex of the New Brunswick Lowlands. The northern boundary is the eastward-flowing Saint John River west of Fredericton, and the eastern boundary is the southward-flowing Saint John River north of Saint John. The St. Croix Highlands are only considered highlands in that they stand elevated above the Bay of Fundy to the south and the New Brunswick Lowlands to the northeast.

The Pokiok Hills, underlain primarily by the Pokiok granite batholith, form the northwestern part of the St. Croix Highlands. They comprise a group of rocky, steep-walled ridges, generally with a northwest-southeast alignment reflecting glacial erosion. Most hillcrests are above 220 m elevation, the highest being at 406 m. Local relief generally is around 100 m. The hills and ridges are well drained, but some narrow valleys contain lakes or are poorly drained because of the glacial diversion of streams away from Saint John River.

The Magaguadavic Highlands form a belt of relatively flat terrain lying between higher granite-cored hills to their south and north (Gadd, 1973). The terrain surface is undulating and generally lies below 210 m elevation (maximum elevation being 241 m). Local relief is variable, generally only a few tens of metres but in places up to 100 m; maximum local relief is 170 m. Ridges are characteristically streamlined and aligned in a northwest-southeast direction, a result of glacial fluting and drumlinization.

Broad depressions filled with glacial drift are commonly imperfectly to poorly drained. The alignment of most ridges and troughs due to intense glacial erosion has resulted in a parallel drainage pattern except in poorly drained, broad depressions where it is deranged. The Magaguadavic and Digdeguash rivers drain most of this area. The northern part of the Magaguadavic Highlands is an undulating upland with many poorly drained depressions and no distinctive alignment of ridges or drainageways. The most dominant feature of this area is a high ridge, trending in a generally northeast direction (Prospect Ridge, Blaney Ridge) and rising to 240 m elevation. This cuesta is a result of Silurian metasediments being more resistant to erosion than the Devonian granites to the west (Seaman, 1982a).

The Scoullar Hills comprise broad imperfectly drained valleys and lowlands, which separate isolated hills, ridges, and clusters of hills. The hills and ridges commonly stand 120 m above the lowlands where relief of only 30 to 40 m is common. Mount Pleasant at 370 m elevation is the highest point. The northern edge of the Scoullar Hills is marked by hilly terrain with local relief of 50 to 100 m (Juvenile Hills). The Scoullar Hills are marked by few lakes and by deranged dendritic drainage; they are drained mainly by tributaries of Oromocto River.

The Nerepis Highlands are separated from the remainder of the St. Croix Highlands by Douglas Valley. This valley, aligned in a northwest-southeast direction, transects the local northeast-southwest structural trends. The southern part of the Nerepis Highlands includes the Nerepis Hills, which are separated from surrounding valleys and lowlands by scarps between 120 and 220 m high. Within the highland itself, relief is commonly 90 m and up to 200 m in places. It has a maximum elevation of 445 m and is marked by a radial drainage pattern. It is spotted with a few small lakes but is well drained due to its high relief. The northern part of the Nerepis Highlands is moderately well drained, undulating, and marked by a dendritic drainage pattern. Ridges and hills generally rise 45 to 75 m above intervening valleys and basins. In the central part, cuestas are present with structurally controlled scarps having 75 m relief. The core of the Nerepis Highlands is drained by Nerepis River, whose lower reaches are incised 120 to 250 m below the upland surface.

The Clarendon Hills are a group of high hills on the eastern part of the Saint George Batholith. They stand well above the Magaguadavic Highlands to the west and the Musquash Lowland to the south. Their northern boundary is formed by a 60 to 90 m-high fault-line scarp. The hills reach a maximum elevation of 330 m with local relief generally about 75 m; rarely local relief exceeds 180 m. Broad rounded hills and ridges separate valleys, which contain numerous lakes due to glacial damming and diversion of drainage. Most of the area drains to the Bay of Fundy via Musquash, Lepreau, New and Pocologan rivers.

The Bocabec Hills are a narrow range of rugged rocky hills, which separate the lowland bordering the Bay of Fundy from the undulating upland surface of the Magaguadavic Highlands. Maximum elevation is just over 200 m and relief is generally between 70 and 100 m. Ridges and valleys commonly have a north-northeast alignment. The area is well drained due to the steepness of hillslopes, even though lakes are common in the valleys.

The Musquash Lowland consists of an undulating plain and isolated clusters of hills bordering the Bay of Fundy. Generally the maximum relief is about 60 m. The hills reach elevations near 90 m, except along the northern edge of the lowland and between Loch Alva and Grand Bay where a few hills rise to over 150 m and relief is up to 75 m. The drainage is deranged due to glacial activity and to the effects of postglacial submergence and emergence. The Musquash Lowland is moderately well drained except for swampy areas located in broad basins scoured in the bedrock. Slopes adjacent to the coast and along its numerous bays are steep. Coastal cliffs are generally between 10 and 30 m in height.

The West Isles are a group of islands separating Passamaquoddy Bay from the Bay of Fundy; Deer Island and Campobello are the two largest. Deer Island and the northern part of Campobello Island are low, but hilly with steep-sided hills rising to elevations just over 90 m. The southern part of Campobello Island is relatively flat and lies below 30 m elevation. Coastal cliffs around the islands generally vary between 20 and 50 m in height.

Grand Manan Island consists of a plateau-like upland with a lowland along its eastern edge. The upland is bounded by steep, 50 to 100 m-high coastal cliffs. It has an undulating surface with broad flat-crested hills whose elevations fall from more than 130 m at the northern end of the island to near 70 m at the south end. The eastern edge of the upland is a scarp whose crest is at about 50 to 60 m elevation and whose base is at about 30 m. The lowland flanking this crest falls gently to the sea. The island is characterized by a dendritic drainage pattern reflecting the general eastward slope. It is well drained except for a few low depressions.

Bedrock Geology Summary

The St. Croix Highlands lie mainly within the Magaguadavic tectonostratigraphic zone (Fig. 2; Fyffe, 1982a). It is underlain by intensely deformed Ordovician, Silurian, and Lower Devonian sedimentary and volcanic rocks, which have been intruded by granitoid rocks of Middle Devonian and Early Carboniferous age (Ruitenberg et al., 1977; Ruitenberg and McCutcheon, 1982). The area has been labelled the Fredericton Synclinorium (Poole and Rodgers, 1972; Fyffe, 1982a) and the Magaguadavic tectonic zone (Fyffe, 1982b). Davies (1977) and Ruitenberg and McCutcheon (1982) have separated the area into a part lying north of the major Fredericton Fault (Magaguadavic tectonic or structural zone) and a part to the south of the fault (St. Croix tectonic or structural zone). The northern part is characterized by Silurian greywackes and slates, whereas the southern part is more complex. "Silurian and Lower Devonian wackes and slates... unconformably overlie Ordovician graphitic and pyritiferous slate, siltstone, quartzite and greywacke" (Ruitenberg and McCutcheon, 1982, p. 133) north of granitoid intrusions. "Silurian marine felsic and mafic volcanic and shallow water clastic sedimentary rocks" and "Lower Devonian mafic and felsic terrestrial volcanics and intercalated fluviatile sedimentary rocks" (Ruitenberg and McCutcheon, 1982, p. 133) are present south of the same intrusions. Rocks in this area were intensely deformed and metamorphosed during the Acadian Orogeny. Major northeast-striking thrust faults, which terminate at northwest-striking wrench faults also transect the rocks in the Magaguadavic tectonostratigraphic zone. Some folding and faulting occurred during the Hercynian Orogeny (Ruitenberg and McCutcheon, 1982).

"The southeast part of the St. Croix Highlands (mainly the Musquash Lowland) lies within the Caledonia tectonostratigraphic zone. Here upper Precambrian sedimentary and volcanic rocks (Green Head and Coldbrook Groups), are overlain in part by Cambrian and Ordovician sedimentary rocks of the Saint John Group and Carboniferous clastic sedimentary rocks" (Ruitenberg et al., 1977, p. 175). West of Saint John River in the western volcanic belt (Ruitenberg et al., 1977), the volcanic rocks are characterized by a mylonite zone and dyke swarm (Ruitenberg and McCutcheon, 1982). Older rocks were deformed and metamorphosed during the Precambrian.

The area has subsequently been deformed during the Taconian, Acadian, and Hercyian orogenies in a fashion somewhat similar to deformation in the adjacent Magaguadavic tectonostratigraphic zone.

A5 – CALEDONIAN HIGHLANDS

Physiography

The Caledonian Highlands are that part of the New Brunswick Highlands (Bostock, 1970) that lies east of Saint John River between the Bay of Fundy to the south and the New Brunswick Lowlands to the north. The Caledonian Highlands have complex geomorphology and geology. These highlands include (1) a central plateau area underlain primarily by Precambrian rocks and (2) boundary areas underlain primarily by folded and faulted Carboniferous rocks and local areas of older rocks (Rampton and Paradis, 1981c). The boundary areas are typically characterized by ridge and valley topography.

The Central Plateau has been described as follows: "A gently rolling plateau-like surface with local clusters of rounded hills characterizes the topography; its surface declines gently at its southwestern end. The plateau is moderately well drained except locally on flat areas and in depressions. The drainage pattern is predominantly radial with streams showing mainly dendritic patterns and being deeply incised at the edge of the plateau surface. At the southwestern end of the plateau the drainage has been glacially deranged and shows an irregular pattern: streams here are less incised than those to the northeast" (Rampton and Paradis, 1981c, p. 2). Maximum elevations of 360 to 400 m are common along its northwestern edge. Local relief is generally only 60 to 90 m except near the coast where streams are incised up to 270 m. Coastal cliffs of 150 m height are common.

Prime examples of ridge and valley topography paralleling structural trends are present in the Anagance Ridges physiographic subdivision, especially its central and western parts. Ridge-top elevations of 215 to 305 m are common in the northeast. Drainage and drainage patterns have been described as follows: "Drainage is generally good in this area of high relief except in some broad valley bottoms clogged by glacial drift or filled with alluvium and intertidal sediments. The drainage has a trellis pattern characteristic of ridge and valley topography. The major streams such as the Petitcodiac, Anagance, Kennebecasis, Millstream, and Hammond Rivers and Smith and Belleisle Creeks parallel the structural trend over most of their course. Tributaries are both obsequent in that they are at right angles to the main structural trend and consequent in that they parallel structural trends. In areas where the ridge and valley topography is not well developed, such as between Petitcodiac and Sackville, glacially deranged drainage patterns reflect regional slopes and range from parallel to dendritic to irregular" (Rampton and Paradis, 1981c, p. 4).

Physiographically, the Moncton Basin is a broadening of one of the valleys within the Anagance Ridges subdivision. Its surface lies below 75 m elevation and is characterized by flat to gently undulating topography and only moderately good to poor drainage. Streams exhibit a deranged drainage pattern because of the area's glacial and postglacial history (Rampton and Paradis, 1981b).

The Shepody Ridges include those ridges and valleys to the east of the Central Plateau formed of folded and faulted Carboniferous strata. In the New Horton area the strongly northeast-southwest aligned ridges reach elevations of more than 90 m, whereas hills and ridges along the edge of the Central Plateau commonly rise to over 150 m; Shepody Mountain has an elevation of 327 m. Stream incision has resulted in 90 to 150 m relief along the edge of the Central Plateau. Coastal cliffs are generally less than 60 m high, but rise to over 180 m high near Alma. Drainage patterns are similar to those characterizing the Anagance Ridges.

The Dorchester Upland consists of two prominent ridges east of Petitcodiac River and Shepody Bay. The ridges are in turn separated by Memramcook River. The southern part of the upland consists of a relatively flat, poorly drained plain standing only 45 to 60 m above sea level. The northern part consists of well drained ridges, which rise to between 120 and 165 m elevation. Streams show an irregular dendritic drainage pattern, even though most ridges appear to be structurally controlled. Intertidal flats forming a lowland adjacent to Memramcook River are included in the Dorchester Upland subdivision.

The Mispec Plateau is an undulating to hilly highland that stands as an intermediate bench between the Central Plateau and the Bay of Fundy. Parts of the boundary between the Mispec and higher Central Plateau are marked by a 45 to 90 m-high discontinuous bedrock and structurally controlled scarp. The surface of the Mispec Plateau generally stands 75 to 150 m above sea level. Local relief is generally below 60 m except where streams have entrenched themselves up to 90 m into the upland surface. Coastal cliffs average about 45 m in height, and rarely they are up to 120 m high. Flat areas on the upland surface are commonly poorly drained. Parallel drainage patterns, which follow regional deformation patterns (Potter et al., 1979; Ruitenberg and McCutcheon, 1982), are dominant on the Mispec Plateau.

Bedrock Geology Summary

The Caledonian Highlands are underlain mainly by pre-Carboniferous deformed rocks within the Caledonia tectonostratigraphic zone (Ruitenberg et al., 1977) and by Carboniferous strata which show a measure of folding and faulting at the surface (Poole and Rodgers, 1972; Fyffe, 1982b) within the Moncton Basin.

In the Caledonia zone (Fig. 2) near Saint John, Precambrian carbonate and clastic rocks of the Green Head Group are the oldest known rocks in New Brunswick (Ruitenberg and McCutcheon, 1982). They are overlain mainly by felsic volcanic rocks of the Coldbrook Group in the western section of the Caledonian Highlands, and mainly mafic volcanic rocks with minor amounts of felsic volcanics in the eastern section (Ruitenberg et al., 1977). In the western part, the Cambrian to Lower Ordovician Saint John Group, consisting of shale, siltstone, sandstone, conglomerate, quartzite, and minor limestone, unconformably overlies the Precambrian strata. "Mississippian and Pennsylvanian fluviatile sedimentary and volcanic rocks also lie upon Precambrian rocks in part of the zone. The youngest rocks in the zone are Triassic red beds and volcanic rocks, which occupy a few small areas along the Bay of Fundy coast" (Ruitenberg and McCutcheon, 1982, p. 133). As reflected by fault-line scarps, the younger rocks are in fault contact with the Precambrian rocks in many places.

The Moncton Basin (Fig. 2) is composed of Carboniferous clastic nonmarine fluviatile and deltaic sedimentary rocks, and red clastics, limestone, gypsum, and salt of the Windsor Group (Poole and Rodgers, 1972; Ruitenberg and McCutcheon, 1982). During the Carboniferous rocks in the Moncton Basin were folded and thrust and block-faulted along fold axes and fault planes with a northeasterly strike (Ruitenberg and McCutcheon, 1982). The upper part of the sequence near Moncton has not been deformed to the same extent (Poole and Rodgers, 1972).

Within the Caledonia tectonostratigraphic zone, the Greenhead Group was deformed and metamorphosed during the Precambrian. On Kingston Peninsula later deformation is commonly difficult to distinguish from Precambrian deformation because of re-activation of faults along older fault zones during later deformation events. Deformation, including thrusting, wrench and block faulting, and arching, was most intense during the Acadian and Hercynian orogenies in the eastern part of the Caledonia zone (Fundy Cataclastic subzone, Ruitenberg and McCutcheon, 1982). The northern part of the Caledonia zone constituting the northern edge of the Central Plateau (Loch Lomond Deformed Zone) is less intensely deformed (Ruitenberg et al., 1979).

A6 – NEW BRUNSWICK LOWLANDS

Physiography

The New Brunswick Lowlands are part of the Maritime Plain, which encompasses much of the Maritime Provinces (Weeks, 1957; Williams et al., 1972). It is a flat to gently sloping plain, which occupies a large triangular area in the eastern and central part of the province. In general, the lowland forms a gently eastward sloping plain, Curventon-Bathurst Valley and Grand Lake Basin being the only interruptions.

The part of the New Brunswick Lowland that lies east of Moncton and contains the arm of land terminating at Cape Tormentine has been called the Tormentine Peninsula. Rampton and Paradis (1981b, p. 4) have described it as follows: "East of the Caledonian Highlands the New Brunswick Lowland is marked by a broad flat area south of Shediac and a gentle ridge, which extends eastward from Scoudouc to Cape Tormentine. The ridge is flanked by gentle slopes and has an indistinct broken crestline with elevations rarely exceeding 40 m (125 ft). Drainage is moderate on most slopes, but is only fair in swales, which have the appearance of drift-filled preglacial valleys. On the broad flat area south of Shediac, drainage is moderate to fair because of its flatness. The streams in this area show an irregular drainage pattern characteristic of glaciated and submerged flat areas. Most streams draining the ridge running from Scoudouc to Cape Tormentine have a north-south orientation."

The Central Lowland constitutes most of the area lying south of Miramichi River and draining into Gulf of St. Lawrence. Its surface rises gently inland from the sea at its eastern edge to elevations of just over 185 m at its western edge where it merges with the Eastern Miramichi Highlands (Rampton and Paradis, 1981a,b). Broad flat areas are separated by poorly defined valleys. Drainage over most of the Central Lowland is "only moderate to fair because of its flatness and because of the Quaternary events, which have affected the area (glaciation and marine submergence). Only in areas where the streams are incised and local relief is up to 75 m (250 ft), such as adjacent to Southwest Miramichi River, near Upper Gaspereau and adjacent to Buctouche River and Cocagne River estuaries, is drainage particularly good. The main streams outline a parallel pattern with a northeast-southwest orientation, which reflects the orientation of the Carboniferous basin and the structural trend within adjacent physiographic units. The lower courses of streams have been drowned during the recent submergence of the region. Tributaries show an irregular drainage pattern because of derangement due to glaciation and marine submergence of the flat Lowland" (Rampton and Paradis, 1981b, p. 2). The coastline is characterized by low scarps of less than 10 m height and many spits and baymouth bars.

The Grand Lake Basin encompasses the land within the New Brunswick Lowland that slopes towards Grand Lake and Oromocto River valley. It generally lies below 120 m elevation. It includes the broad flat valleys occupied by Oromocto and Saint John rivers, large expanses of these valleys being swampy. Saint John River transects the basin and drains it to the south. The upland surface is flat to gently sloping with the flat areas being poorly drained. Only in the area of Camp Gagetown and between Grand Lake and Washademoak do well drained rolling hills occur; relief is generally below 45 m, even though a maximum elevation of 140 m is reached. Incision by small streams in the Grand Lake Basin locally has resulted in relief of 45 m. Saint John River is incised 90 m below the adjacent upland surface in a few localities.

The Rusagonis Hills form a rolling hilly upland along the western edge of the New Brunswick Lowland to the south of Saint John River. The hilly nature of the area is largely due to the incision of the Saint John and its tributaries. Upland relief is only about 15 m but increases near streams (Lee, 1957; Seaman and Thibault, 1981). Ridges and hills commonly crest near 210 m elevation within a few kilometres of valley axes. The most prominent ridge of the area is the cuesta to the south and west of Oromocto Lake, which is formed by a competent eastward-dipping Pennsylvanian sandstone overlying a recessive erodible Carboniferous unit (Seaman, 1982a). The steep edge of this cuesta forms an escarpment up to 100 m high. The Rusagonis Hills are well drained except for a few poorly drained broad depressions north of Oromocto Lake. Drainage patterns range from dendritic in the northern part to deranged in the southern part.

The Nashwaak Hills are that part of the New Brunswick Lowland north of the Saint John River where stream incision of 150 m below the general upland surface by Nashwaak River and its tributaries has given the landscape a hilly nature. Even though Nashwaak River tributaries are incised up to 100 m, the relief on the undulating upland surface exceeds 60 m. The area is well drained because of the high relief. Major tributaries of the Nashwaak have developed a parallel drainage system at right angles to the river (Rampton and Paradis, 1981a).

The Curventon-Bathurst Valley is a north-south oriented trough that interrupts the gently eastward-sloping surface of the Eastern Miramichi Highlands and Acadian Peninsula. Only the base of the trough is included in the Curventon-Bathurst Valley subdivision as the upper edge of the trough is locally difficult to define, and bedrock to the west of the trough axis is typical of that underlying the adjacent Miramichi Highlands. The base of the trough is even difficult to define at its northern end because of the gradual rise away from the trough axis. Scarps up to 90 m high border the trough at its southern end. The valley bottom is characterized by flat to gently sloping terrain with imperfect to poor drainage. It merges imperceptibly at its northern end with the West Bathurst Basin. The valley is occupied by the lower segments of Nepisiguit and Northwest Miramichi rivers and by Portage River and Gordon Meadow Brook.

The Acadian Peninsula is basically a large-scale cuesta. The land surface slopes gently upward from the Gulf of St. Lawrence more than 150 m elevation and then descends abruptly to the base of the Curventon-Bathurst Valley in a series of escarpments and benches. The eastward-sloping surface has a gently undulating form except where dissected. Isolated groups of bedrock-cored hills, such as Butte à Morrison, rise 30 m above the general surface. Eastward and southward draining rivers, such as the Pokemouche, Tracadie, Tabusintac, and Bartibog, are incised between 30 and 80 m

below the adjacent uplands (Gauthier, 1983). Relief of 5 to 30 m is even present along their rias. The eastern coastline is characterized by spits and baymouth bars, whereas sea cliffs up to 30 m high are present along the northern coast. Drainage along incised streams and rias is moderately good, but drainage on broad flat areas near the headwaters of streams is poor and characterized by deranged drainage patterns (Gauthier, 1983). Drainage is also poor in broad low flat areas near sea level and in the Rivière Caraquet drainage basin, which is characterized by a subparallel drainage pattern due to its glacial history. Drainage on the westward sloping scarps and benches bordering the Curventon-Bathurst Valley is variable according to the steepness of the slope. Some benches are poorly drained.

The Acadian Islands, consisting of Miscou, Shippegan, and Pokesudie islands, are an eastern extension of the Acadian Peninsula. The islands are relatively flat and poorly drained with extensive peat bogs. Maximum elevation is just over 23 m. The coastlines are characterized by scarps less than 5 m high, spits, and baymouth bars.

Bedrock Geology Summary

The New Brunswick Lowlands are underlain primarily by Carboniferous rocks that characterize the Central (or New Brunswick) Platform (Fig. 2; Poole and Rodgers, 1972; Fyffe, 1982a). They consist primarily of Pennsylvanian sandstone, but locally siltstones or shales, conglomerates, and coal measures are present; they reach a maximum thickness of 1000 m near the coast (Poole and Rodgers, 1972). Isolated outcrops of Mississippian and Silurian rocks are present in the southern part of the lowlands (Potter et al., 1979) and a diabase dyke of Triassic age transects the Acadian Peninsula (Hamilton and Gupta, 1972; Gauthier, 1983). South and east of Bathurst, the Bathurst Formation, consisting predominantly of redbeds, can be differentiated from the Clifton Formation consisting of greenish and brownish sandstone with a few interbedded red shales (Davies, 1977; Gauthier, 1983). The latter is more typical of the Pennsylvanian rocks underlying the remainder of the New Brunswick Lowlands (cf. Fyffe, 1982a). The Pennsylvanian strata dip gently eastward except near the Caledonian Highlands where they are gently folded (Poole and Rodgers, 1972).